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HYDROLOGY AND SEDIMENT TRANSPORT IN THE ELBOW RIVER BASIN, S.W. ALBERTA

by

© H.R.HUDSON

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled HYDROLOGY AND SEDIMENT TRANSPORT IN THE ELBOW RIVER BASIN, S.W. ALBERTA submitted by H.R.HUDSON in partial fulfilment of the requirements for the degree of Ph.D..

DEDICATION

This thesis is dedicated to my wife Trudy.

ABSTRACT

The 1210 km^2 Elbow River basin, which drains the mountains, foothills and plains of the Eastern Slopes of the Rocky Mountains, was studied to investigate the hypothesis that runoff and sediment are derived unevenly in time and space from complex, intermediate sized, drainage basins. A network of 18 primary measurement stations was operated for two years to supplement available hydrologic and sediment yield data. Long term rates of suspended sediment, bedload and solute load transport were estimated from measured discharge - load relationships. Sediment sources and controlling processes were identified based on aerial photographs and field measurements.

The sediment yield regime of the Elbow River basin is controlled by the hydrologic regime and the sediment supply regime. The steep, gravel bed river has a typical high latitude runoff regime. Flows are low during winter when the river is frozen over. Winter snow fall is effectively redistributed so that it is largely removed over a short period to produce a spring snow melt flood which may be enhanced by rainfall. About 60% of the total annual discharge occurs during the three month snow melt period. Summer rainfall may generate several, generally smaller, flood peaks following the main spring flood. Small magnitude floods tend to be generated from snow melt in the mountains and upper foothills. Rainfall may complicate the response. Large floods are produced as the result of a combination of

rainfall and/or snow melt from the whole of the basin.

Reservoir surveys show that the average suspended sediment output of the Elbow River system, since dam construction in 1932, is about 75600 t/yr, or 62.5 $t/km^2/yr$. The mountain suspended sediment yields, which average 2100 t/yr, or 11 $t/km^2/yr$, are derived largely from surface wash of barren, calcareous colluvium, within tributary basins. Yield is limited by sediment supply. Weathering of the calcareous bedrock produces little fine material and the finer textured valley wall morainic deposits are usually protected by forest or lag deposits at their toe. The majority of the annual transport occurs in spring when the ephemeral tributaries flow and when mass movements into the stream channel tend to occur. The upper foothills sub - basin yield of 3950 t/yr, or 16 $t/km^2/yr$, is derived almost equally from tributary and river channel and valley wall erosion. The lower foothills sub - basin yield of 12150 t/yr, or 34 $t/km^2/yr$, is derived almost exclusively from along the river channel. Upland erosion of the potentially highly erodible foothills is limited by vegetation protection and/or sediment trapping in tributary log jams and beaver dams. Bank and riparian erosion is also the primary source of the 57400 t/yr, or 138 $t/km^2/yr$, derived from the lower basin. Upland erosion of the potentially highly erodible glaciolacustrine and till surficial materials of the plains may occur as the result of rare runoff events.

On the rising limb of the annual spring flood there is a strong relationship between discharge and suspended sediment concentration. Following the peak discharge, a rapid drop in sediment concentration occurs, with a marked shift in the suspended sediment rating curve. About 55 to 75% of the total annual suspended sediment load is transported in ten days and 98% is transported during the high flow period (May, June and July). These features are explained in a conceptual bank and valley wall erosion model.

Bedload has been measured at several sites in the Elbow River basin using large basket and Helley - Smith bedload samplers. Significant time and spatial variations in the measured rates of bedload transport are attributed to sediment supply constraints, the mechanics of bed material motion and a downstream decrease in competence. Criteria for evaluating bedload formulae are proposed and several formulae are tested against measured loads at four sites on the Elbow River. Some frequently used formulae were totally unsuitable, whereas other could be modified to adequately describe the sustainable capacity bedload. Calibrated formulae were used to estimate the sustainable bedload capacity (i.e. the amount of bedload transport which would occur if there were no sediment supply - constraints) for the four main hydrometric stations on the river. The average mean capacity load increases downstream from the mountains through the foothills, but decreases through the plains

(mountains 783 t/yr, upper foothills 2335 t/yr, mid foothills 13453 t/yr, lower basin 1013 t/yr). The average annual load, which reflects supply constraints, is approximately one quarter of the mean capacity load. The downstream discontinuity in bedload transport is accommodated by storage as sediment waves in the generally degrading channel.

In the mountains surface wash processes and throughflow readily dissolve the predominantly calcareous bedrock and regolith. The foothills are a groundwater recharge area. Rapid throughflow and saturated overland flow may inhibit solution of calcareous and sedimentary rock and regolith. In the plains the solute load is derived primarily from transient, regional groundwater flow influxes into the river valley, with very limited tributary and local groundwater inputs. Solute yields average 98 $t/km^2/yr$ in the mountains, 65 $t/km^2/yr$ in the upper foothills, 54 $t/km^2/yr$ in the lower foothills, and 25 $t/km^2/yr$ in the plains.

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1. INTRODUCTION

A. RATIONALE

Fluvial processes are the dominant contemporary sediment transfer processes which result in the erosion, transport and deposition of sediment at the earth's surface (Gregory and Walling, 1973). This continuing redistribution of sediment has resulted in numerous problems (American Society of Civil Engineering, 1975).

Most sediment problems are related to one or more of three aspects of sedimentation:

1. erosion may cause a deterioration of land surface quality and quantity;
2. transportation of sediment may cause physical and aesthetic damage to water quality; and
3. sediment deposition may effect specific land and water uses (Guy, 1970; Painter, 1976).

"The extensiveness and severity of sedimentation problems has made sedimentation a vital concern in the conservation, development and utilization of ... soil and water resources" (ASCE, 1975:2). The redistribution of sediment is a natural process which may be accelerated or alleviated by the action of man. Because there are generally few benefits resulting from sediment redistribution, such as floodplain replenishment, the general objectives of sedimentation engineering are to design for sedimentation processes, or to reduce natural erosion and minimize accelerated

erosion, so that related sediment transport and deposition problems are reduced.

The resolution of a sediment yield problem has four main steps: (1) the nature of the problem has to be recognised and understood, (2) the sources of sediment must be determined and (3) controlling processes must be identified so that (4) control measures may be evaluated and implemented. For example, reservoir sedimentation problems are generally caused by the total sediment load, whereas water use problems are generally caused by excessive fine materials carried in suspension, and channel problems are generally more a function of the bed material.

Once the nature of the problem is recognised and the critical characteristics of the sediment are established, the sources of the problem sediment and the processes controlling sediment supply must be determined. For example, sediment may be derived almost exclusively from upland erosion sources, such as rilling and gullying (Glymph, 1954), whereas in other areas stream bank erosion (Hansen, 1971) or mass movements into the stream channel (Fredrickson, 1970) may be the major sediment source. The detachment and transport processes are obviously quite different and require unique control measures. In most cases it would be physically or economically impractical to treat every sediment source, hence the relative rates of contribution, and the paths and modes of transport should be determined to maximize the utility of control measures.

B. SEDIMENT YIELD RESEARCH APPROACHES

Sediment yield research may be divided into studies of: (a) the upland, or out-of-channel phase, (b) the in-channel phase and, (c) combinations of the upland and channel phases. In this section examples of each approach, and the limitations of previous research, are examined.

Upland erosion studies

The four major sediment transfer processes (surface, sub-surface, channel, and mass erosion) may be subdivided into numerous discrete and interactive processes. Each process requires particular measurement techniques (Gregory and Walling, 1973:145-151).

The ultimate aims of upland erosion measurement programs are usually to: (a) describe and quantify the rate of processes, (b) rationally account for process operation, (c) to predict patterns and rates at one site, and (d) to extrapolate these findings to other sites. However, there are a number of problems. For example, Hayward (1967) outlines some of the difficulties of making physically meaningful measurements in soil erosion plot studies. Further, because of the complexity of the system it may be difficult to rationally account for processes. For instance, Bryan (1968), examined various indices of soil erodibility and found that none had universal application. Extrapolation to other areas and conditions, therefore, may be tenuous.

Even in the ideal situation where measured erosion rates are accurate, and rationally accounted for, the extrapolation to other sites may yield gross erosion rates which bear little relationship to the actual export of sediment from the drainage basin, because yield is not synonymous with erosion (ASCE, 1975), at least in the short term (Wilson, 1973). Sediment yield is a reflection of the gross erosion in, and the efficiency of sediment removal from, a drainage basin (Roehl, 1962).

Erosion from a site may be related to sediment yield to establish a sediment delivery ratio. This ratio can be applied to estimated gross erosion in a drainage basin to predict sediment yield (Renfro, 1975). However, a number of factors influence sediment delivery from an erosion site (Renfro, 1975; Meyer et al., 1975). This makes the application of a general sediment delivery ratio, usually based exclusively on drainage basin size (Renfro, 1975), quite unrealistic (Boyce, 1975). Boyce (1975) suggests the application of differing sediment delivery ratios for sub-areas of a drainage basin delineated by slope. This would be an improvement, but what is required are sediment delivery ratios for each erosion site, and these may vary with time (Dickinson and Wall, 1977).

Although there are a number of drawbacks which limit the extrapolation of on-site measurements, the approach may ascertain the relative importance of various sediment sources. Further, when used in conjunction with stream flow

sampling, the approach may provide the means for understanding denudation rates and patterns, because consideration is given to both the upland phase and the channel phase.

<u>The in-channel phase</u>

Three widely used methods quantify sediment transport by stream flow: (a) direct measurement of sediment in motion by various sediment sampling devices, (b) direct measurement of sediment accumulation in some type of settling pond and, (c) load prediction by formulae.

<u>(a)Direct measurement of sediment in motion:</u>

This method requires concurrent field measurement of stream flow and sediment load. If sediment concentration is known for relatively small increments of the stream flow hydrograph, then an integration of instantaneous sediment discharge will produce the sediment transport for any given period. The frequency of sampling depends on the character of a given drainage basin. If discharge - concentration relations are erratic, then very frequent samples or continuous measurement are required. Often some synthesis is necessary. The most common approach is to establish an average relationship between stream flow and sediment concentration. Sediment concentration is measured over a wide range of flow conditions to establish a sediment rating curve for a particular stream. The sediment rating curve is then combined with stream flow duration data to compute sediment yield by correlation (Miller, 1951).

Although the sediment rating curve approach is applicable to bedload transport (Strand, 1975), the approach is normally limited to suspended sediment transport, because of the difficulty of measuring bedload. The sediment rating curve approach is inherently weak because it relates sediment yield to discharge which is only one of the contributing factors (Glymph, 1954). Kellerhals et al., (1972) address this problem in some rivers in Alberta. They found, for instance, that in the lower reaches of the Red Deer River local rainfall over the badlands may produce large influxes of sediment without an appreciable increase in the discharge of the river.

There are numerous problems regarding field and laboratory techniques (Einstein, 1947; Hubbell, 1964; Douglas, 1971; Loughran, 1971), sample frequency for load calculation (Walling, 1977b), and the method of calculating loads (Meade, 1969), and delineation of non-denudation load components (Singh, 1970). Although direct measurement methods are probably the most costly and time consuming, they may permit insight into erosion processes which are impossible to gain by other methods (Piest et al., 1975).

(b) The sediment accumulation method:

The sediment accumulation survey method involves periodic surveys to estimate the volume of sediment deposited in reservoirs and other settling ponds. Alternatively, sediment cores have been dated using 137 cesium (Ritchie and McHenry, 1973). The volume of sediment trapped is converted

into a weight, adjusted for trap efficiency, and expressed as a rate of accumulation (hence hinterland sediment yield) beween surveys (Glymph, 1975). Bed load and suspended load are usually not differentiated by the method (Gregory and Walling, 1973). In some studies traps have been resurveyed during significant flows to establish the rate of accumulation (Hollingshead, 1968; Nanson, 1974). Generally, however, the survey period is seasonal (Anderson and Wallis, 1965), or longer (Hollingshead, 1969), so that while this method provides general information on the long term magnitude of yield, it lacks details on the timing of inputs and sources of material (Agricultural Research Service, 1975). The major advantage of the method is that it provides long term sediment yield records comparatively cheaply (Glymph, 1954).

(c) Load prediction by formulae:

Sediment transport equations usually consider motion of individual rock fragments of various sizes and ignore material in solution (ASCE, 1975). These formulae describe two distinct types of motion which follow different laws: traction, whereby particles move near the bed such that their weight is supported primarily by the bed; and suspension, whereby particles are suspended and supported by the stream flow. This division is consecrated by conventional techniques of measuring sediment in motion (Einstein, 1947). It is recognised that the division between suspension and traction is arbitrary and dependent on flow competence. Gravel and larger material, however, usually moves along the

stream bed whereas material finer than sand usually moves in suspension (Simons and Senturk, 1977). The sand fraction (0.063 to 2.00 mm) is dynamic and may move in either or both modes simultaneously (Church and Gilbert, 1975).

The load which is carried in suspension is divisible into suspended bed material load and wash load (Simons and Senturk, 1977). The former consists of material found in appreciable quantities in the active stream bed and wash load is defined as finer material which forms less than ten percent of the bed material deposits. This finer material is supplied from river bank erosion and from upland erosion. The term "finer material" is relative. Simons and Senturk (1977) suggest that wash load consists of the silt and clay fractions in a sand bed stream and, that sand, silt and clay could be considered as wash load in coarse gravel and cobble bed streams.

Theoretically, there should be a well defined functional relationship between the force exerted on the stream bed and the quantity of bed material moved. The mode of transport, traction or suspension, would depend on flow competence. Numerous formulae have been developed (ASCE, 1975). However, there are few data which can be used to evaluate bed material transport, particularly in gravel bed streams (Leopold and Emmett, 1976). Various evaluations of these formulae suggest that, in field conditions, predicted transport rates may bear little correspondence to measured loads (ASCE, 1975; Church, 1976; McLean, 1980). This may be

due partially to the selection of inappropriate formulae. The selection of an appropriate formula is hindered because: "Comprehensive guidelines for selection and application of appropriate sediment-transport formula for use in natural stream channels do not exist." (ASCE Task Committee, 1982:1353).

Wash load, by definition, is sediment supply-limited and therefore can not be predicted in a similar manner to the bed material load. However, there may be a symptomatic relationship between wash load and stream competence in that conditions which promote bed material transport may also provide upland sediment to the river channel (Bogardi, 1972). As a result wash load is usually related to selected drainage basin indices using regression techniques (e.g. Anderson, 1967; McPherson, 1975).

Combined upland erosion and in-channel approaches

The conceptual framework for the interaction of upland erosion and in-channel erosion was described by Leopold, Wolman and Miller (1964). However, Walling (1977a) suggests that few studies of upland erosion, in conjunction with in-channel measurement, have been attempted. He cites the work of Slaymaker (1972) and Imeson (1974). In fact there are a number of other studies which consider relevant aspects of the two phases together. The substantial literature on river bank erosion often assesses the contribution of bank erosion to the total sediment load (e.g. Hansen,1971;

Carson et al., 1975). The literature on sediment delivery ratios is based on measurements or estimates of gross erosion and sediment yield from basins or river reaches (e.g. Maner, 1958; Roehl, 1962; Renfro, 1975). In addition, non-quantitative estimates of upland erosion have frequently been made in inventory studies. Neill and Mollard (1982), for example, photographed major sediment contributors in the Oldman River basin, Alberta. In this, and other inventory studies, (e.g. Brown and Jackson, 1974, in California) the connections between individual erosion sites and the channel network are not elucidated. These types of studies can be extended with considerable field work and examination of sequential air photographs to quantify the contribution of individual sites to the stream sediment budget (e.g. Kelsey, 1980, in the Van Duzen River basin, north coastal California).

Quantitative analysis of sediment sources and sediment yields suggests that the relative importance of the major erosion processes (surface, subsurface, mass movements and stream channel erosion) varies considerably with geographic location and land use. In the sparsely vegetated arid S.W. United States, for example, surface wash processes are the dominant component of sediment yield (Glymph, 1954). In badlands of Alberta, surface wash processes produce very high yields from limited areas within the Red Deer River basin (Campbell, 1977). Conversely, in well vegetated alpine basins stream channel erosion is the almost exclusive source

of sediment yield. In addition, mass movements may be locally important in this environment and vegetation removal may lead to greatly increased sediment yields from surface wash processes (Swanston and Dyrness, 1973; Fredricksen, 1970; Megahan, 1976).

The combined approaches to sediment yield studies suggest that suspended sediment yield in well vegetated basins is usually derived from small portions of the drainage basin as the result of stream channel and riparian erosion. Limited observations suggest that bedload is derived from channel erosion and from upland erosion (Fredrickson, 1970; Nanson, 1974). Recent research indicates that solute loads are also unevenly derived in time and space from a drainage basin (Hughes and Edwards, 1977; Walling, 1977b).

Conclusions

Sediment yield studies generally treat the contributing drainage basin as a "lumped system" in that they deal with the basin as a whole, rather than with its constituent parts, in deriving the quantity of sediment expected at a point downstream (Glymph, 1975). In the "lumped" approach sediment yields are related to gross basin characteristics such as area, slope and drainage density (e.g. McPherson, 1975). Sediment yields from these studies are mathematically redistributed back over the drainage basin to estimate an average rate of erosion. Although this approach has provided a broad regional picture of sediment yield (e.g. for Canada,

Stichling, 1973), "relatively little attention has been directed to the study of causative processes which produce and supply sediment to streams" (Slaymaker and McPherson, 1973:170). Hence, although yield can be related to gross basin parameters, "there is a very large gap in our ability to relate the sediment load in a river to the erosion and transport mechanisms upstream" (Ketcheson, et al, 1973:186). "It follows, then, that we are not nearly as well prepared as needed for making definitive statements about the impacts of our program measures upon sediment yield". (Glymph, 1975:3).

The distributive system approach, where the sources and rates of transport from individual sites are examined, has revealed many shortcomings in the "lumped" approach. However, the distributive sediment yield approach is still in its infancy and much is to be learned, particularly for regions such as the Canadian Cordillera (Slaymaker, 1972). Reviews of the state of the art suggest that a number of fundamental questions should be addressed. Questions of particular importance include: (a) which erosion sites provide what sediment reaching downstream points? (Glymph, 1975), (b) what are the modes, paths, and time constraints of sediment movement? (Wolman, 1977), (c) what role do transient or unsteady phenomena of erosion and transportation play? (Wolman, 1977), (d) what are the relevent climatic and physiographic characteristics which determine sediment transport? (Weyman, 1975; Dickinson and Wall, 1977).

C. THESIS APPROACH

Study area selection

The 1210 km^2 Elbow River basin, which is located in South Western Alberta (Figure 1.1), was chosen as a study area for several reasons. The river system experiences sedimentation problems. Glenmore Reservoir, located in the lower reaches of the Elbow River 12 km above the confluence with the Bow River, is filling with sediment. This is resulting in the loss of recreational facilities and a decrease in water storage capacity at a time of great, and rapidly increasing, demand for both uses. The continuing redistribution of sediment into the reservoir necessitates sediment filtering for municipal water use. Water hardness is also a problem (Meyboom, 1961). Flooding is perceived to be a major problem and flood - sediment control structures have been proposed (Monenco, 1980).

In addition to these practical needs, there are several academic reasons for studying sediment yield in the Elbow River basin. Little is known regarding sediment yield processes in the Canadian Cordillera (Slaymaker, 1972). Hence, although details of the sediment yield and hydrology of the Elbow River basin may be unique, the mountains, foothills and plains of the basin present a microcosm of the Eastern Slopes of the Cordillera (Johnson, 1977) and the principles learned in the Elbow River system study may apply quite widely. Further, most research tends to concentrate on

Figure 1.1 Location of the Elbow River basin

either very small or very large drainage basins, at the expense of intermediate - size systems such as the Elbow River (McPherson, 1975). As well, the Elbow River is one of only a few rivers in Alberta with reasonable background information and reasonable access. The Elbow has been gauged at several points over a relatively long period and a number of meteorological stations are in the area. Also, relatively long term, detailed, suspended sediment data and three years of bedload data are available from the mid point of the system. Glenmore Reservoir sedimentation surveys are also available to provide a long term check on sediment load prediction in the lower basin (Hollingshead, 1969). Finally, various reports, theses and papers, which describe the basin geology, geomorphology, soils, climate, vegetation, land use, and sediment yield, are available.

Study plan

The basic academic objective of this thesis is to evaluate the hypothesis that sediment yield is unevenly derived, spatially and temporally, from a complex, intermediate - size, drainage basin. Further, sediment yield processes are examined in the basin to identify the controls of the sediment regime. To these ends the sources of sediment, and the volume, timing and modes of transport from individual sites, river reaches, sub - basin areas and physiographic divisions (mountains, foothills and plains) are described. These academic objectives are complementary

to applied sedimentation research requirements in the Elbow River basin.

Sources of sediment are examined at four levels: (1) At a basin - scale aerial photographs were used to provide a reconnaissance survey of sediment sources, relative rates of sediment contribution and types of sediment involved. Sedimentation in the Glenmore Reservoir provides a basin - scale measure of the rates, volume and type of sediment being exported from the system; (2) At a sub - basin scale sediment loads are estimated for individual tributaries, based on measurements of load and stream flow and by modelling; (3) River reaches are isolated by input - output analyses of sediment and stream flow; and, (4) At individual sites, identified by the aerial photograph analysis and field observation, rates and processes of erosion from river banks and riparian erosion are described. Details of the measurement programs and results are described in following chapters.

There are four main sections, seven chapters and three appendices in this thesis. The first section is introductory and places the work in the context of other research. The second section describes properties of the drainage basin which are thought to be important for the hydrology and sediment yields of the study area. The third section describes dynamic characteristics of the drainage basin and is presented in sub-sections describing hydrology and sediment transport. The sediment transport section is split into

three chapters which describe suspended sediment load, bed-load and solute load transport. In each chapter the findings are discussed with respect to other research. The final chapter synthesises the major conclusions of the previous chapters. The appendices discuss various aspects of the research in greater detail than is appropriate in the main text.

2. ELBOW RIVER BASIN CHARACTERISTICS

A. INTRODUCTION

This chapter is divided into two main sections: an overview of the environmental setting, and an evaluation of the interaction of the various basin characteristics which together determine the availability and character of sediment for fluvial sedimentation. Energy and mass transfer characteristics of the study area are discussed in subsequent sections dealing with climate, hydrology, hydraulics and sediment transport.

The environmental setting may be considered non - varying for the purposes of this study (Schumm and Lichty, 1965). Basin features are considered to determine hydrologic and sediment yield response (Gregory and Walling, 1973). It is recognised, however, that present basin features, which are largely inherited from glacial and fluvioglacial regimes, are dynamic. Thus, contemporary processes are determined by basin characteristics and in turn the basin is shaped by contemporary processes.

B. BASIN CHARACTERISTICS

The Elbow River flows about 120 km from the Front Ranges of the Rocky Mountains, through foothills and plains into the Bow River at Calgary (Figure 2.1). The general topographic form of NNW - SSE trending ridges and valleys, which decrease dramatically in elevation from the mountains

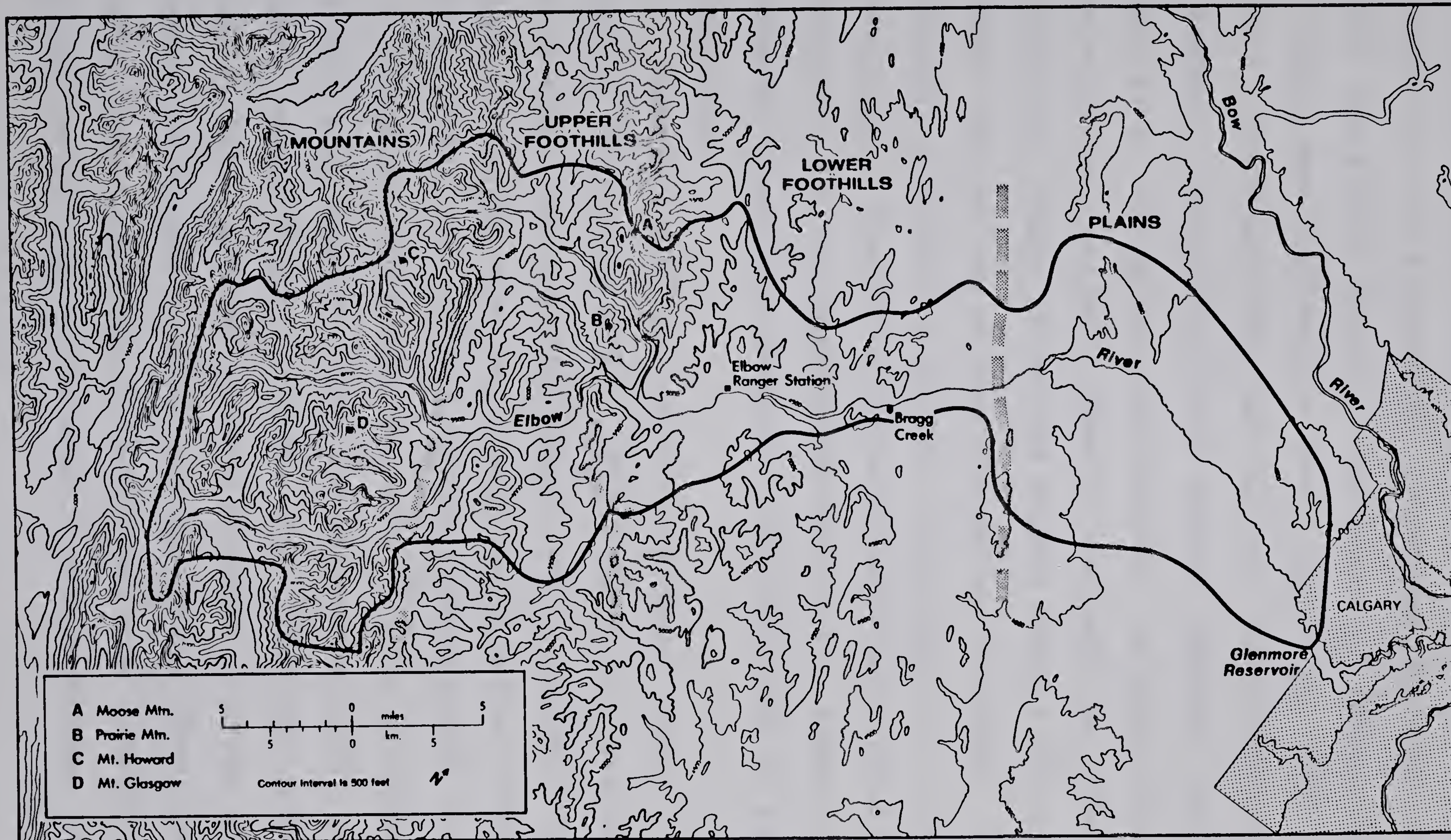

Figure 2.1 Elbow River basin topography and physiographic zones.

in the west to the undulating plains in the east (Figures 2.2 and 2.3), is structurally determined (Evers and Thorpe, 1975). The plains "syncline" represents the eastern extent of thrust sheet deformation (Evers and Thorpe, 1975). The form of individual ridges and valleys is attributed to subsequent glacial and fluvial erosion (Jackson, 1977). The Elbow River and its upstream branches, the Upper Elbow to the south and the Little Elbow to the north, and some major tributaries (Bragg, Canyon and Prairie creeks) flow at almost right-angles to the structual and lithologic trend of the basin (Figures 2.4 and 2.5). (See Figure 3.1 for basin hydrography). Glaciation did not substantially alter the preglacial drainage pattern (Jackson, 1977). Seagel (1971) suggested discordance resulted from antecedence or superimposition.

The Elbow River is one of the steepest rivers in Alberta (Kellerhals et al., 1972). The river flows about 120 km from the Opal and Misty Ranges, which are about 3050 m high, to the Bow River at Calgary, which has an elevation of about 1000 m. The steepest channel gradients tend to be in the mountains with a rapid decrease in slope downstream (Figure 2.2). However, there are exceptions to this in the upper reaches (Figure 2.6). The river flows over predominantly gravel and cobble sized alluvium. Although cross - channel variations in grain size may exceed downstream variations in grain size, bed material from geomorphically similar units tends to decrease in size downstream (Figure

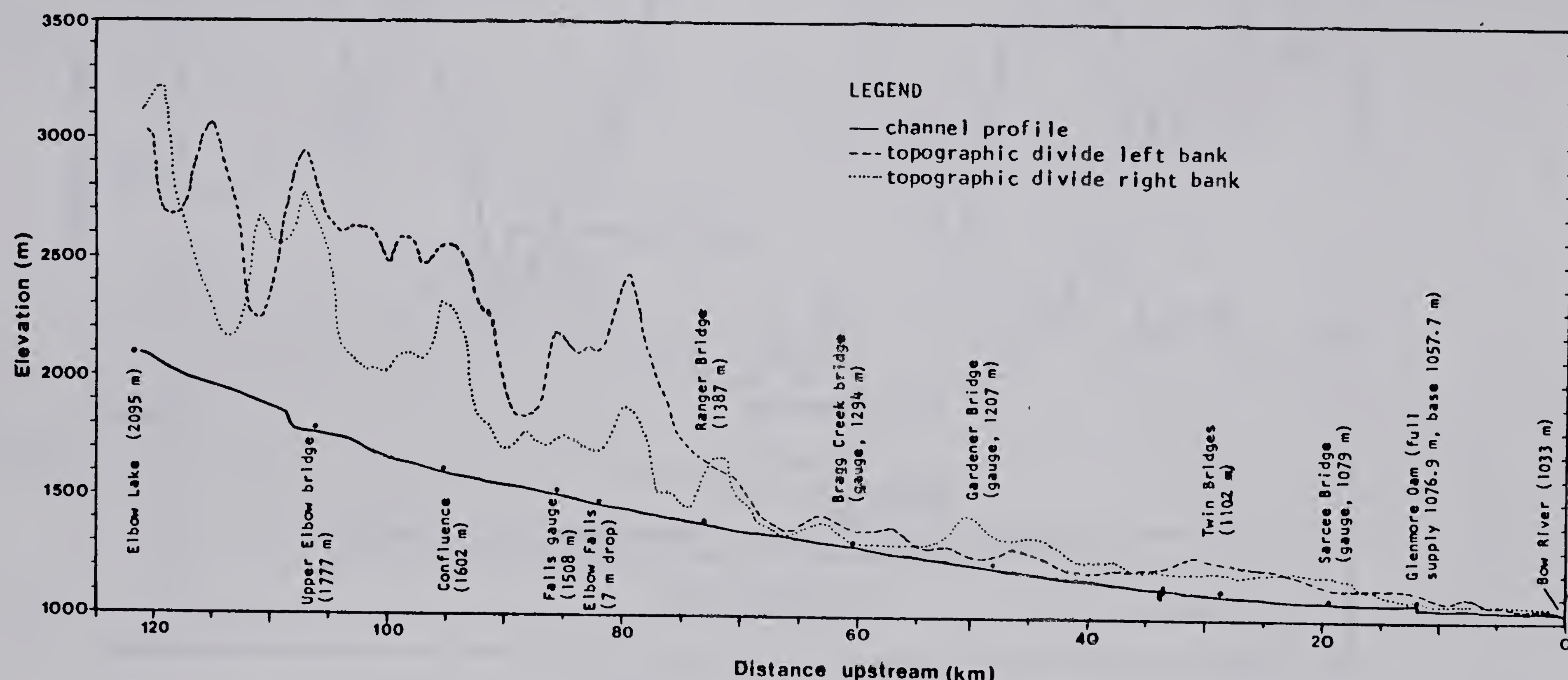

Figure 2.2 Elbow River long profile and watershed divide heights

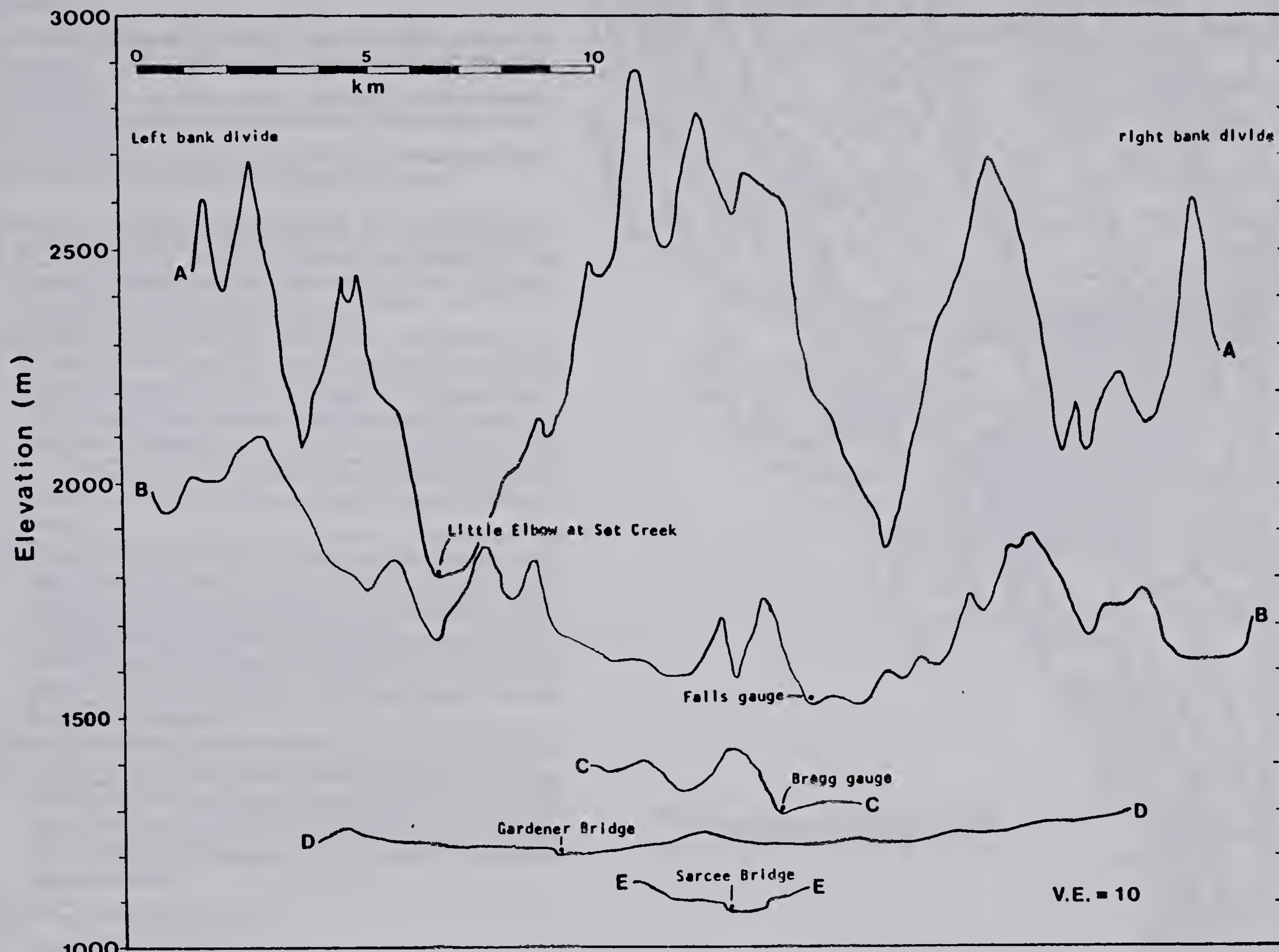

Figure 2.3 Elbow River basin topographic cross - sections

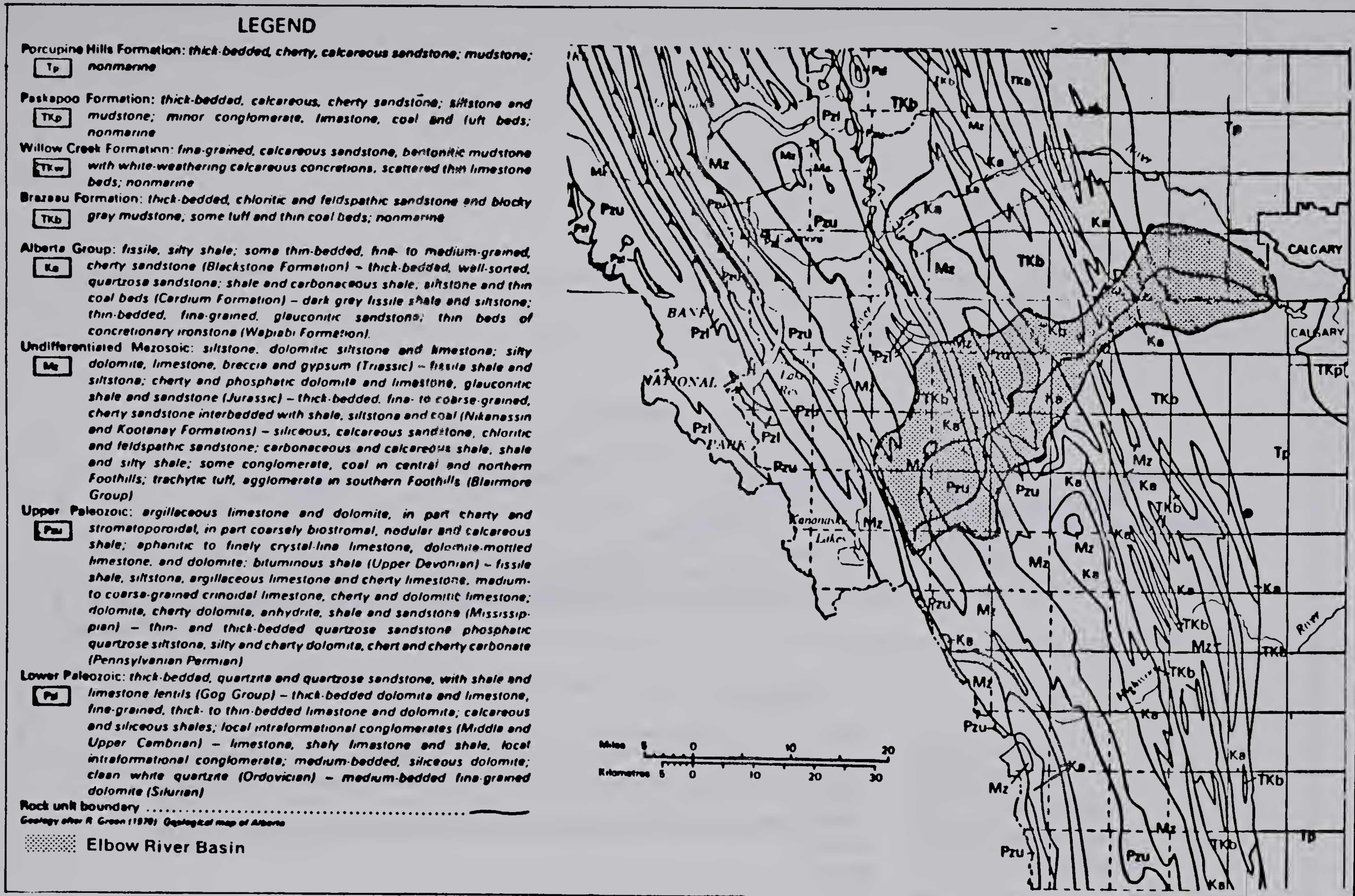

Figure 2.4 Elbow River basin bedrock geology

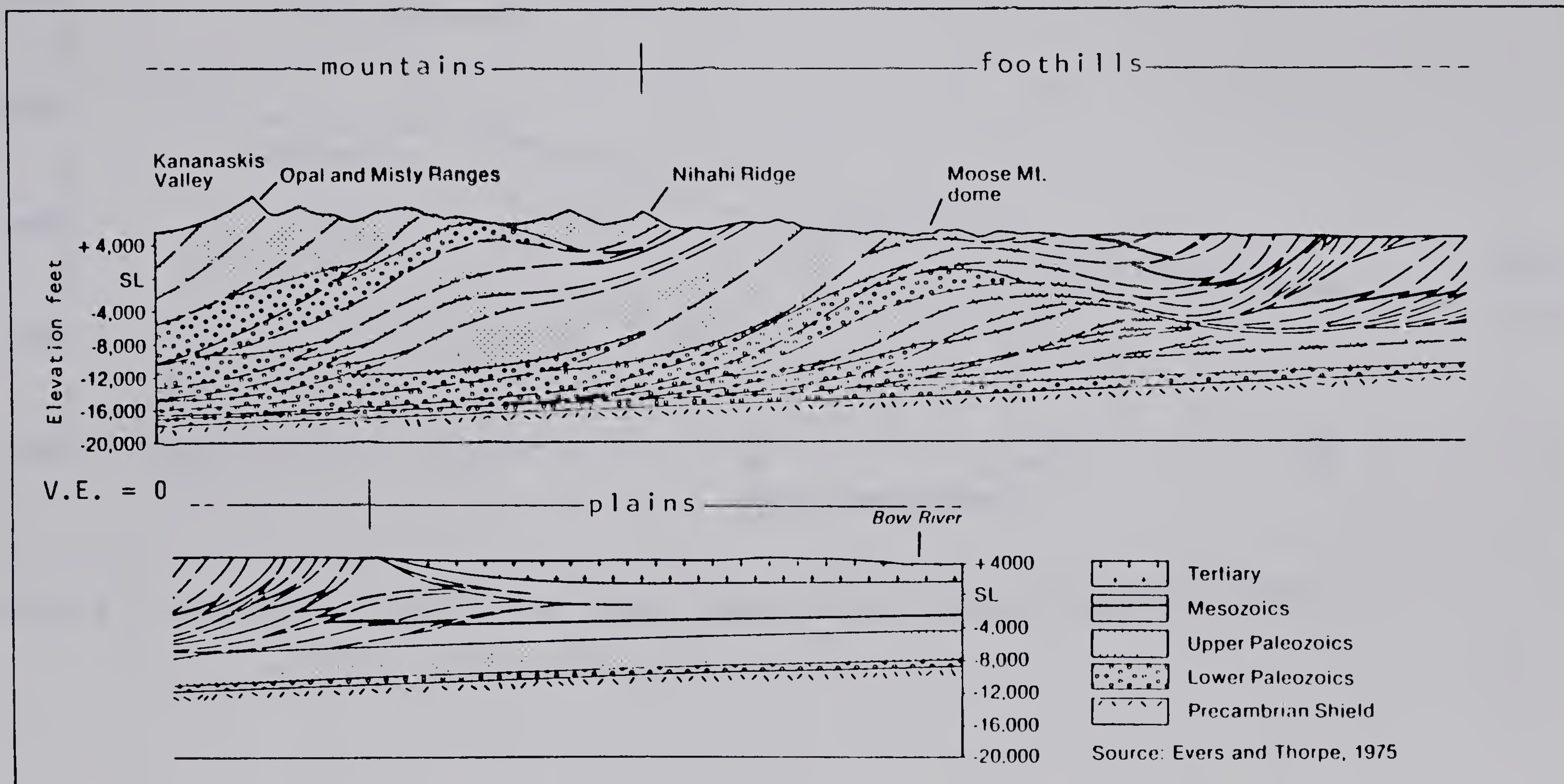

Figure 2.5 Elbow River basin structural cross - sections

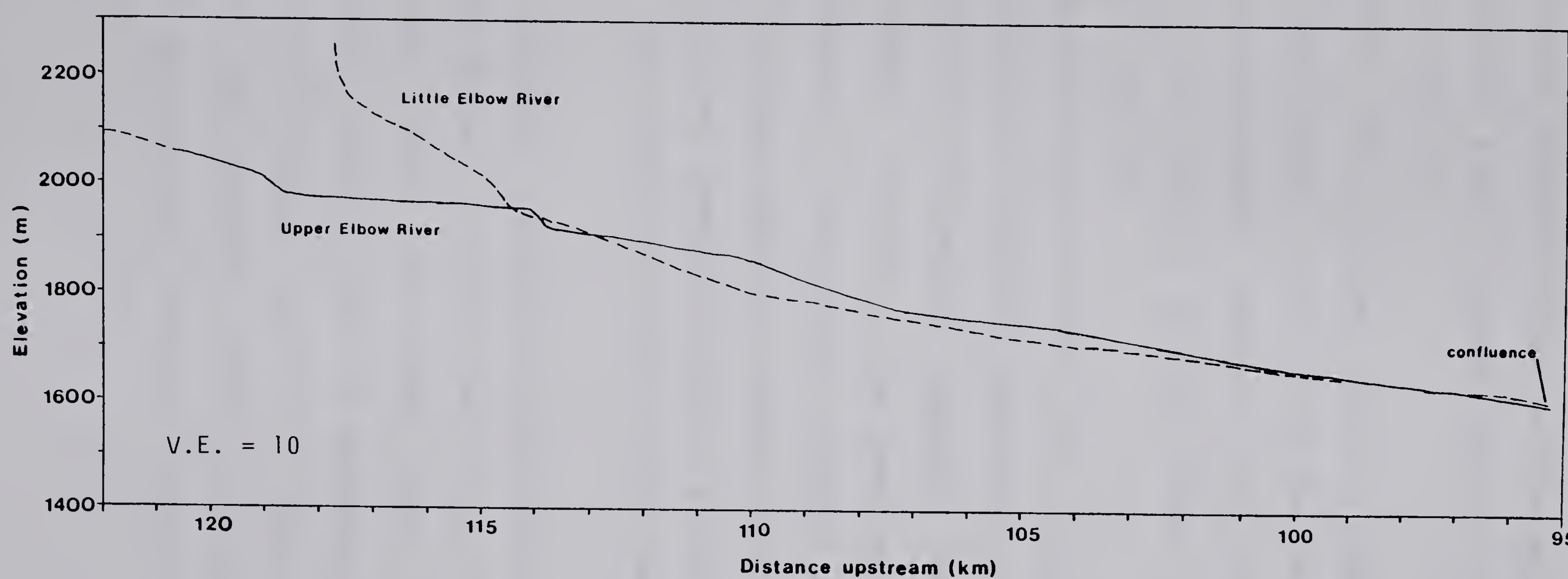

Figure 2.6 Long profile of the Upper Elbow River and Little Elbow River

2.7). The river generally flows in a single channel with a pool and riffle sequence except at the mountain front, where braiding occurs, and in limited bedrock canyon reaches (Photos 2.1 and 2.2).

The Elbow River drainage basin is composed of three structural provinces: (1) the Front Ranges and (2) the Foothills of the eastern margin of the Cordillera and (3) the Plains of the Alberta syncline (Evers and Thorpe, 1975) (Figures 2.4 and 2.5).

The mountains and foothills are underlain by sedimentary strata originally deposited in marine or lagoon environments on a Precambrian basement. These sediments were thrust northeast, relative to the passive basement, to produce a series of imbricated, northwest striking, generally southwest dipping, concave upward, locally folded and faulted thrust sheets (Figures 2.4 and 2.5; North and Henderson, 1954; Monger and Peck, 1972; Wheeler et al., 1972).

Older Paleozoic carbonate rocks were thrust eastward over younger Mesozoic clastic rocks in the Front Ranges (Wheeler et al., 1972). The thrust sheets are often steep to near - vertical at their point of emergence. The Paleozoic rocks tend to form peaks, cliffs and ridges while the recessive or less resistant clastics (shales and sandstones) of Mesozoic age are more frequent in the lower slopes and valleys (Milus et al., 1976).

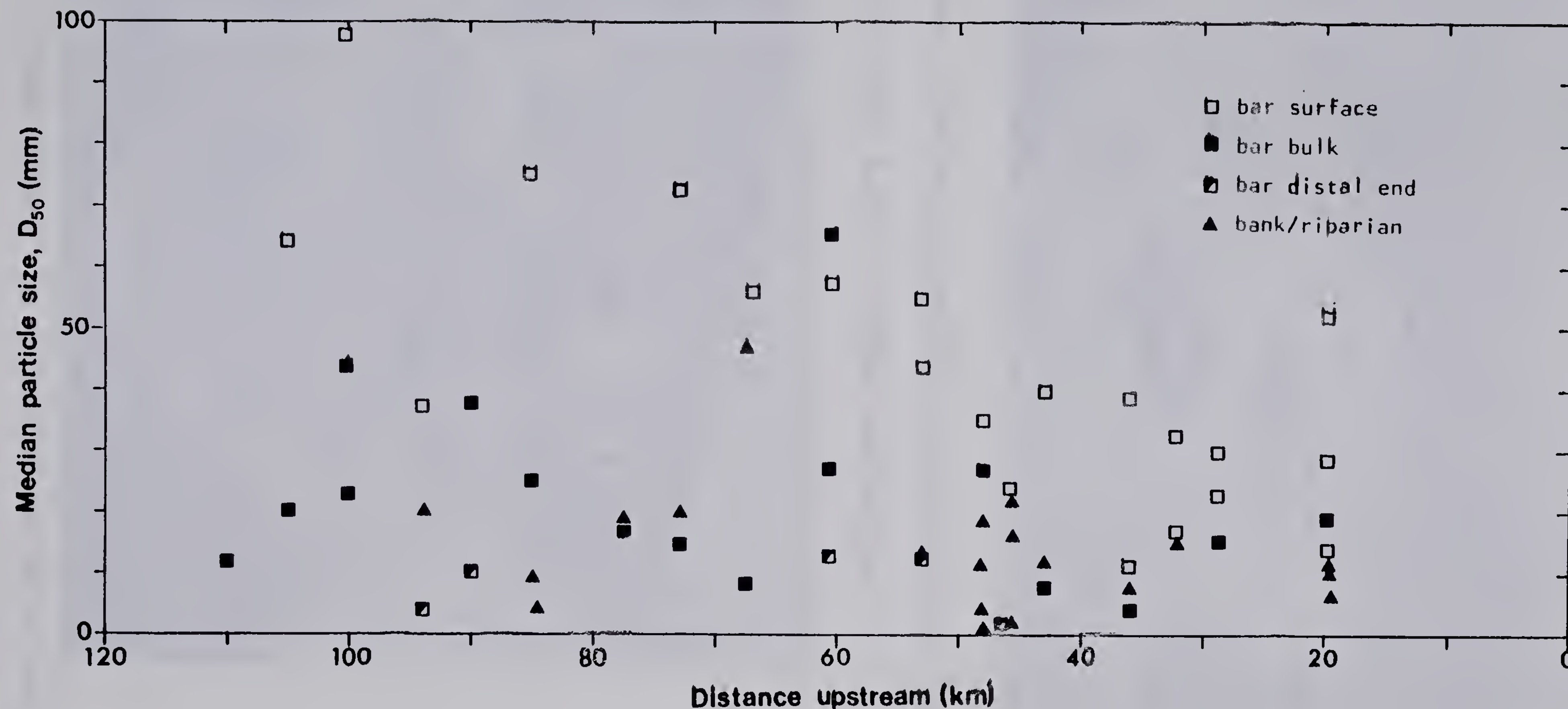

Figure 2.7 Downstream variations in bed material size

Photo 2.1 River channel aggradation occurs in the Elbow River valley at the base of the mountains

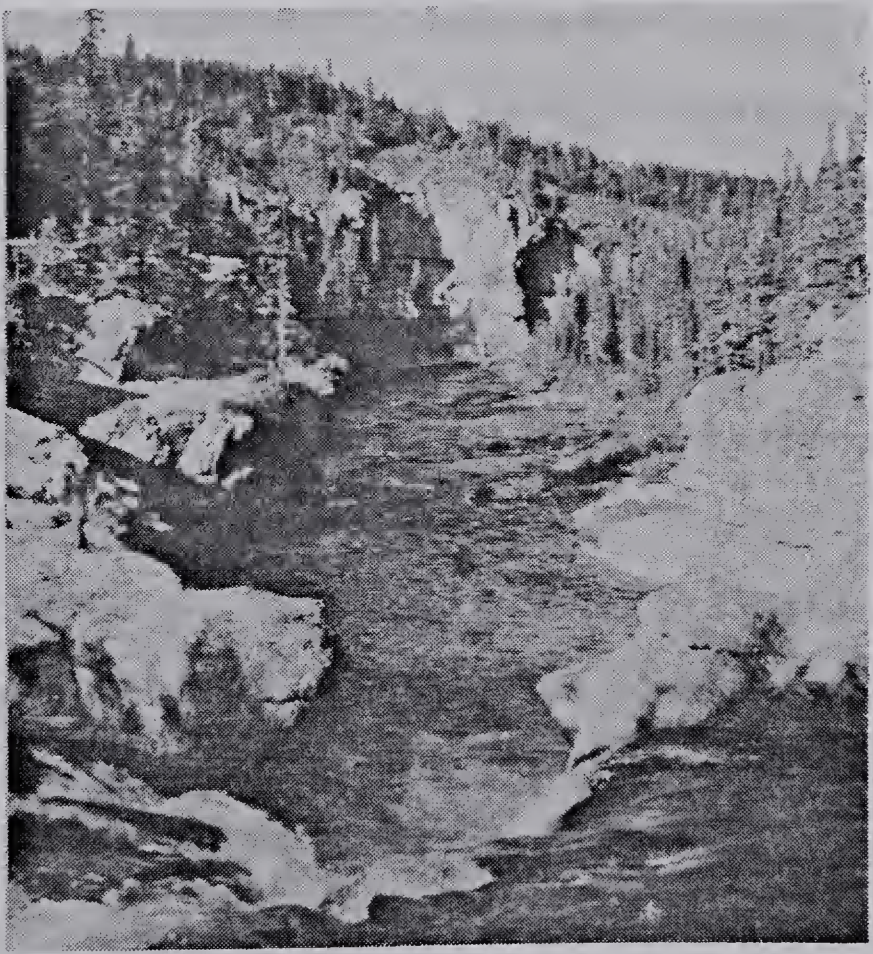

Photo 2.2 The Elbow River is entrenched in a bedrock canyon near Elbow Falls

The McConnell thrust fault divides the Rocky Mountain Front Ranges from the Rocky Mountain Foothills. In the western portion of the foothills thrust plates include Paleozoic strata, whereas in the eastern portion of the foothills thrusting generally affected only the Mesozoic rock that consists of sandstones and shales which lie unconformably on Paleozoic Mississipian strata (Figures 2.4 and 2.5).

The mountains, and the upper foothills west of the Falls gauge area (Figure 2.1), have structurally determined, high relief, ridge and valley topography. The valleys are broad and U-shaped with numerous other glacial erosion features, such as cirques, horns and aretes (Seagel, 1971; Jackson, 1977) (Photo 2.3). Pre - Wisconsin glaciations undoubtedly contributed to the present landscape. However, subsequent erosion and burial have obliterated evidence of these early glaciations in the Elbow River basin (Jackson, 1977; Smith, 1979). The most extensive materials in the mountains are bedrock and colluvium (Figure 2.8). The colluvial materials are derived from limestone and dolomite which produce predominantly gravel - sized fragments with miniscule quantities of finer material (Photo 2.4). These materials have a low surface erosion potential (Table 2.1) and exhibit various slope stability problems (Table 2.2). The bedrock units are prone to failure by mechanical weathering and mass movements.

Table 2.1 Surface erosion potential of barren surficial material units of the Eastern Slopes of Alberta

Surficial material unit	Slope class 0-14%	15-29%	30-44%	45% +
Alluvium (8-14, 21-23)	1	1	2	2
Till (3-7)	1	2	2	2
Colluvium (24)[a]	1	2	2	2
Colluvium (24)[b]	3	3	3	3
Glaciolacustrine (15)	1	3	3	3

Unit numbers refer to Figure 2.8
Unit 24[a] predominantly limestone and dolomite parent material
Unit 24[b] predominantly shale and sandstone parent material

A rating of 1 suggests a low erosion potential, with 0-20% probability of minor rill and gully development.

2 represent an intermediate erosion potential, with 20-50% probability of rill and gully development.

3 represents a severe erosion potential, with a probability of greater than 50% of large rill and gully development.

(Based on Reimchen and Bayrock, 1977).

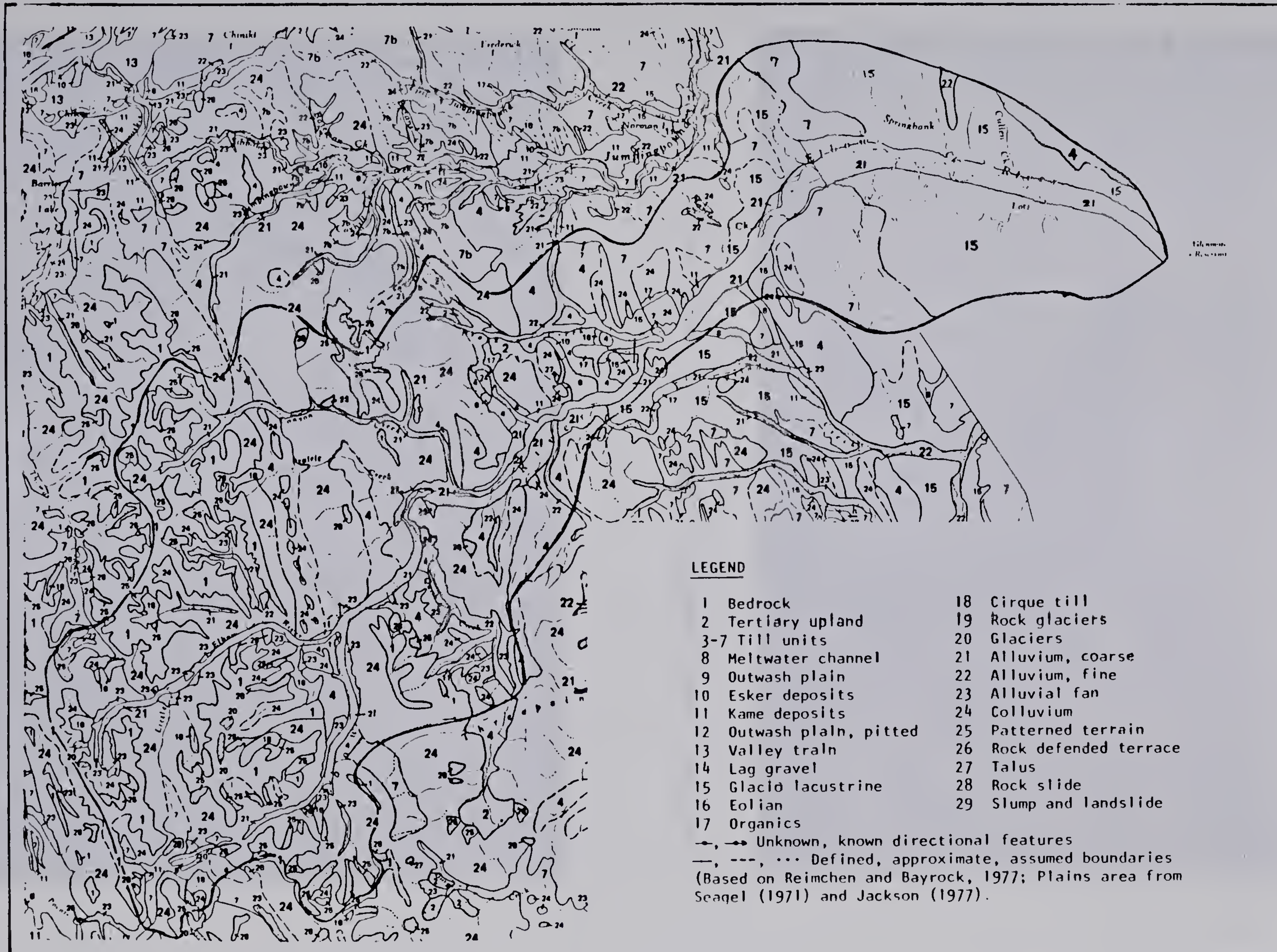

Figure 2.8 Elbow River basin surficial geology

Photo 2.3 View down the Upper Elbow River valley

Photo 2.4 Coarse textured Paleozoic colluvium

Table 2.2 Stability of surficial material units in theEastern Slopes of the Rocky Mountains

Surficial Deposits	Stability			
	Slope Class			
	0-14%	15-29%	30-44%	45% +
Cirque Till (18) Laurentide Till (4) Outwash (8 to 14)	1	1	1	2
Alluvial Gravels (21) Alluvial Fan (23) Rock Slide (28) Patterned Ground (25)	1	1	2	2
Cordilleran Till (3,4,5,7) Rock Glacier (19) Colluvium (24)[a] Talus (27)	1	2	2	2
Glaciolacustrine (15) Eolian Sand (16) Alluvial Sand (22) Slump, Landslide (29)	1	3	3	3
Organics (17)	2	3	3	3
Colluvium (24)[b]	3	3	3	3

[a]Limestone and dolomite and, [b]sandstone and shale parent material.

A rating of 1 suggests the deposits have little or no slope stability problems, except by stream undercutting.

2 represents metastable slopes and 3 represents deposits with high or extreme slope stability problems.

(Based on Reimchen and Bayrock, 1977).

Till and alluvial material cover most of the remainder of the mountains of the Elbow River basin. The tills were deposited mainly from the Early Wisconsin Erratics Train Glaciation, Big Rock Stade (Jackson, 1977). Front Range facies of the Bow Valley Till, derived from Highwood Pass and Elbow River Basin glaciers, are found in the mountains. The subsequent Canmore Advance of the Erratics Train Glaciation reworked much of these deposits. As a result the Canmore Tills (map unit 7 in the mountains) have similar characteristics to adjacent Bow Valley Tills (map unit 4 in the mountains) (Jackson, 1977). Limited glacier advances in the last stade of the Erratics Train Glaciation, the Eisenhower Junction Advance, deposited local cirque tills (map unit 18). The stability and erodibility of the various till deposits are summarized in Tables 2.1 and 2.2.

Alluvial deposits in the mountains of the Elbow River Basin consist of predominantly gravel, stable, low erodibility alluvial fans (map unit 23) and fluvial and glaciofluvial plains and terraces (map unit 21), which are restricted to the valley bottoms.

In the upper foothills bedrock outcrops are far less prominent than in the mountains, but colluvium remains the dominant surficial material cover (Figure 2.8) (Photo 2.5). The major till in this area is Bow Valley Till, which was derived from local cirque and Front Range glaciers (Jackson, 1977). The following Canmore and Eisenhower Junction Advances were restricted to minor cirque glacier advances in

Photo 2.5 View of the foothills from Moose Mountain

Photo 2.6 Extensive alluvial deposits occur in the river valley of the foothills

the mountain outliers of the foothills (Jackson, 1977). Major alluvial deposits occur in the valley bottoms (map units 21 to 23, and 26; Figure 2.8) (Photo 2.6).

Further eastward, in the mid and lower foothills, glacial erosion appeared to be limited mainly to truncation of spurs along the Elbow River valley (Seagel, 1971). Here the foothills are covered with extensive, relatively deep, colluvial and morainic deposits. The ridge tops are often covered with colluvial blankets which grade downslope into morainic blankets. The valleys are usually infilled with glaciolacustrine deposits or alluvium. These deposits are often covered by organic materials. The tills in this area grade from Front Range facies of the Bow Valley Till in the west to mixed facies Bow Valley Till in the Bragg Creek area (Figure 2.8). The mixed facies till was derived from Main Range glaciers which moved down the Bow River Valley and swept southward into the Elbow River Basin where the glaciers coalesced and mixed with Erratics Train Till ice from the Athabasca Valley. The Athabasca ice tongue moved to the study area from the north, approximately 300 km, as a "buffer" on the western edge of the south - west moving Laurentide ice sheet (Jackson, 1977). The mixed facies till grades into Erratics Train Till in the lower foothills (Jackson, 1977).

Colluvial development from recessive sandstones and shales on ridges in the foothills may be interglacial or periglacial, because ice either did not over - ride the

ridges, or scour was negligible (Laycock, 1957). The shale colluvium is highly erodible and unstable (Table 2.1).

In the lower foothills and plains potentially highly erodible and unstable (Tables 2.1 and 2.2) Sheep River Silts and Clays were deposited in glacial lakes. These glacio-lacustrine materials (map unit 15) in the lower foothills were deposited in Glacial Lake Bragg during the retreat of Bow Valley and Elbow Valley ice at the end of the Big Rock Stade of the Erratics Train Glaciation. The retreat of the Bow Valley ice acted, in association with uplands in the Bragg Creek area, to dam the Elbow valley producing Glacial Lake Bragg (Jackson, 1977). This lake later drained southwards and formed other lakes at lower elevations further to the east. Glacial Lake Calgary formed in the Bow River Valley and deposited Sheep River Silts and Clays (map unit 15) over Erratics Train Till (map unit 7) on the plains during the last phases of the Big Rock Stade. These surficial material deposits, which overlie gently folded Tertiary sandstones and shales, are generally a few to several meters thick (Borneuf, 1980). Bedrock outcrops are limited to a few exposures on the plains and in the Elbow River valley.

After Glacial Lake Calgary drained, the Elbow River probably began flowing in its present course, rather than down the Priddis Creek meltwater channel (Jackson, 1977; Figure 2.8). The preglacial course of the lower Elbow River is, however, controversial (Jackson, 1980). Thus, the

outwash plains (map unit 11) and alluvium (unit 21), which form the wide floodplain in the lower foothills and plains of the Elbow River valley (Figure 2.8), were largely derived during the last deglaciation phases of the Big Rock Stade (Photo 2.7). Subsequent Canmore and Eisenhower Junction glaciations, did not directly affect the lower foothills and plains, but probably contributed to alluvial redistribution during deglaciation. Terraces and hillslopes were mantled with loess containing Mazama Ash. The ash is sandwiched within eolian silts which mirror the underlying topography. This suggests that "Terrace cutting and much of the erosion in most areas of the easternmost foothills and prairies was nearly completed by about 6600 B.P. (the time of the Mazama ash fall...)" (Jackson, 1977: 188).

Seagel (1971:127) suggested that at present "a quasi - equilibrium may have been attained, with a greater tendency for degradation" in the river channel upstream of Bragg Creek. Downstream of Bragg Creek "the coarse gravel reaches are being degraded as finer gravels are carried downstream..." which produces a marked decrease in grain size due to aggradation from near Twin Bridges downstream into the reservoir (Seagel, 1971: 129, 131). Hollingshead (1969:18), however, states "There is no evidence that aggradation has worked upstream further than the full supply level of the reservoir". Recent river channel stability is discussed in a later chapter as these contrary suggestions have very important implications with respect to the

Photo 2.7 View of Elbow River valley in the lower basin

Photo 2.8 View of the plains of the lower Elbow River basin

sediment budget of the Elbow basin.

Vegetation is altitudinally zoned in the Elbow River basin. The plains area naturally supports prairie grasslands and woodlands, which have been largely cleared for pasture, cultivation, and acreage development, except in the river valley and on valley walls (Photo 2.8). Downstream of Glenmore Reservoir the basin has been urbanized. Upstream the valley wall vegetation consists mainly of white spruce, with some balsam poplar woodland associations on moister north and east facing slopes. Grass an poplard aspen groves are supported on the drier slopes. The floodplain vegetation changes from mainly balsam poplar near the reservoir to predominantly white spruce upstream.

The lower foothills support aspen poplar on the valley walls and white spruce on the floodplain. The middle and upper foothills vegetation is composed largely of aspen poplar and willow scrub on the floodplain, with pine and/or spruce on the river terraces and hill slopes. The montane zone, which extends from the foothills to tree line (about 1900 to 2300m), consists of coniferous forest. About 65 percent of this forest is seral lodgepole pine, which grew following extensive burning during railroad construction in the Bow Valley in the 1880's (Johnson, 1977). The remainder of the forest is mainly climax spruce. The alpine tundra, which extends from tree line to the alpine barrens, is devoid of large trees, but has a well developed herbaceous mat.

The major land uses of the foothills and mountains are as a forest reserve with an emphasis on recreation, grazing, and petroleum exploitation. The most significant disturbances to the vegetation cover result from fires, limited logging mainly in the Bragg Creek basin, plus roads and seismic cut lines. Recent fires have been contained within small areas (Figure 2.9). Numerous seismic lines have been cut in straight lines, irrespective of topography and stream crossings, throughout the Elbow River basin. Johnson (1977) suggested that although many of these trails are old enough to have revegetated, the use of off - road vehicles is preventing a return to natural conditions, particularly in the foothills.

C. SEDIMENT SUPPLY

An interpretation and mapping scheme was developed by the writer (Hudson, 1977) in which the type of erosion processes, relative rates of action, size characteristics of the erosion product and relative rate of delivery to a stream network are determined for individual erosion sites, at a basin - scale, using aerial photographs. The basic tenet of the approach is that erosion processes produce distinctive "fingerprints" which are observable, or may be inferred to occur, on aerial photographs (Figure 2.10 and Table 2.3). Given an understanding of how the processes operate, the resultant fingerprints may be used to infer sediment supply to a stream network. The Elbow River basin

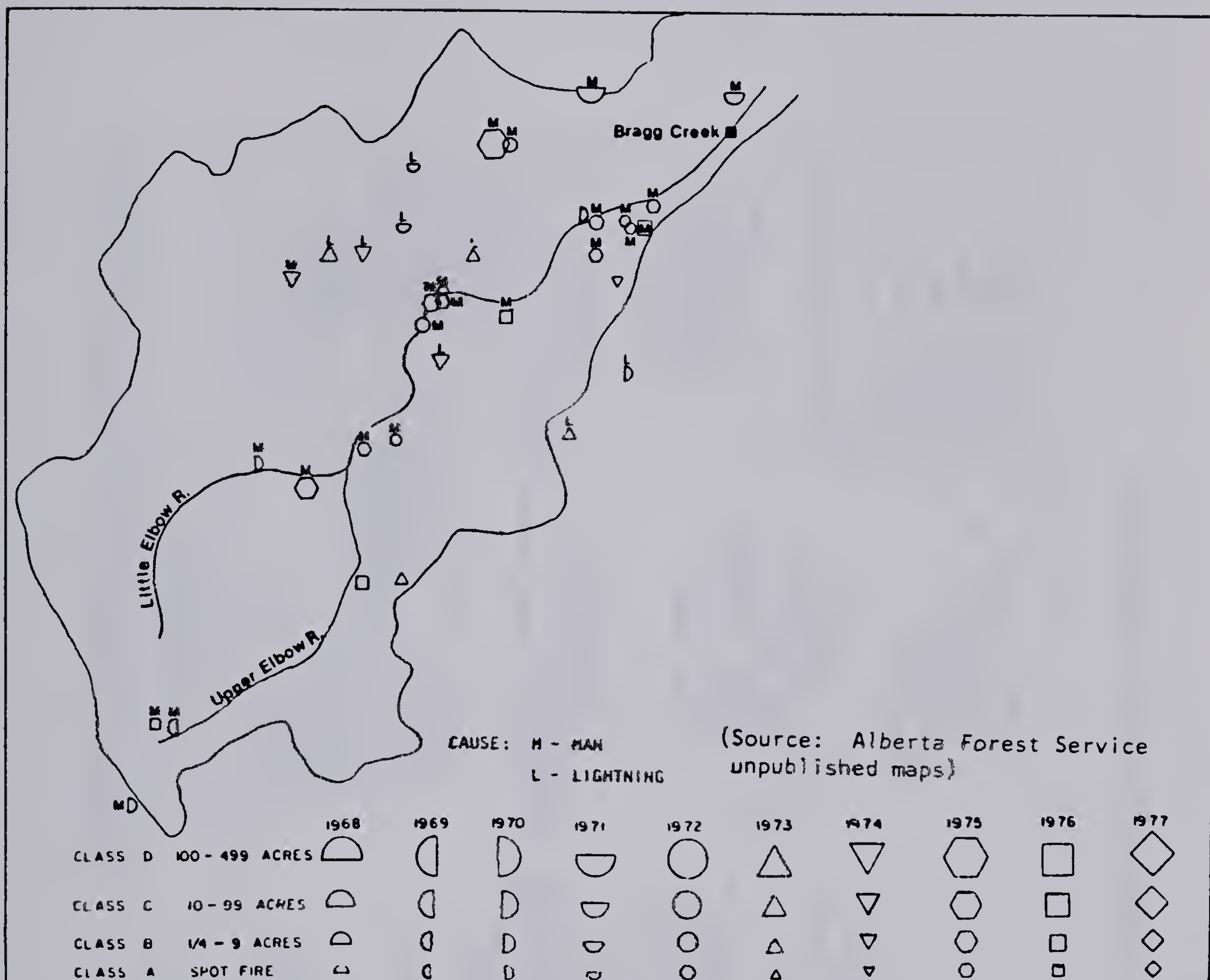

Figure 2.9 Recent fire history of the foothills and mountains of the Elbow River basin

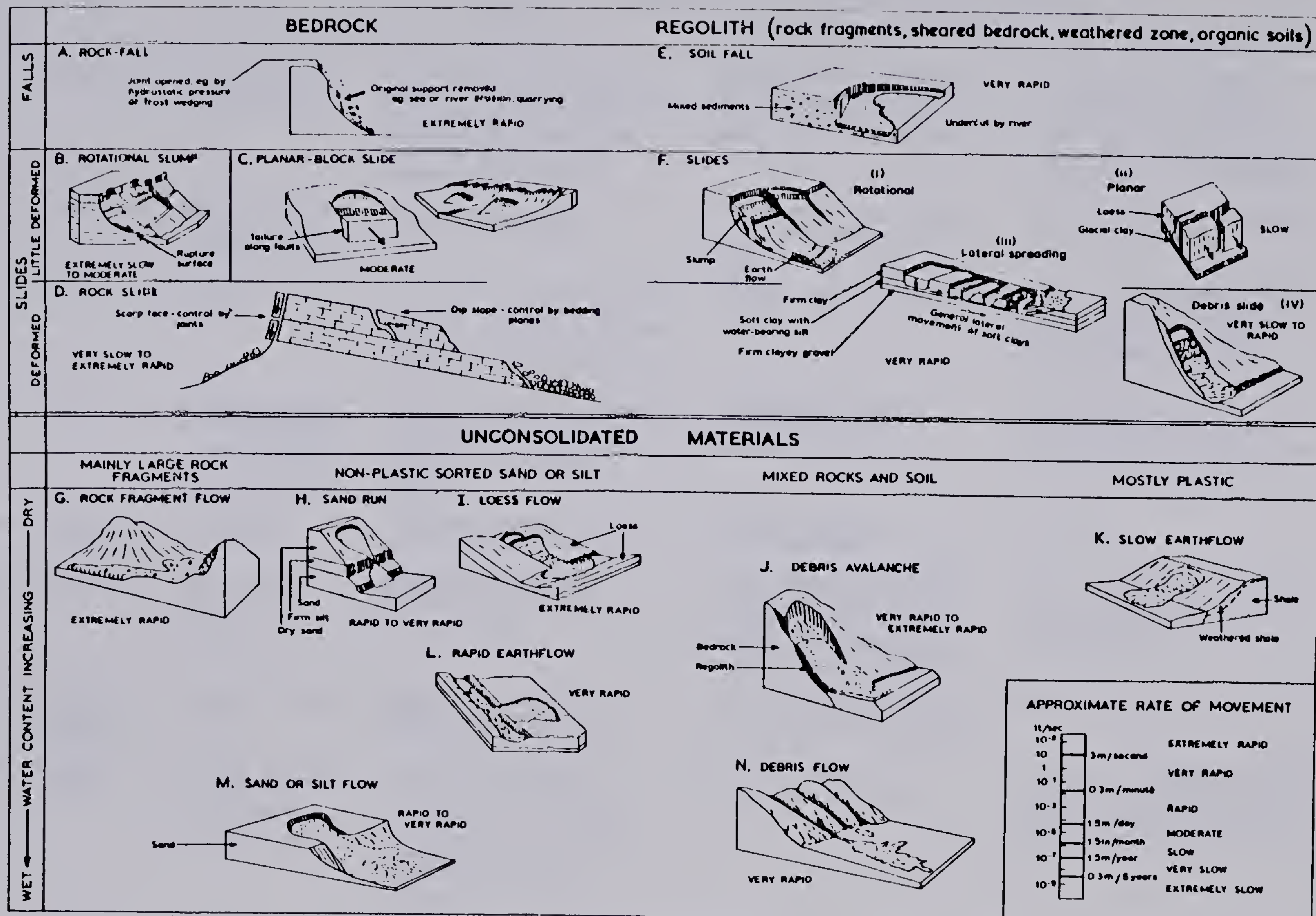

Figure 2.10 Mass movement identification features

(Based on Varnes, 1978)

Table 2.3 Criterion for the identification of sediment supply areas using aerial photographs

Symbol	Type of erosion	Process	Location	Recognition of resultant erosional forms: field	aerial photos	problems	Erosion product
Sp	surface	rainsplash	bare or sparsely vegetated areas	truncation of soil profile, exposure of roots, removal of fine material, coarse material lag, clogging of soil pores with fine material, soil splashed on vegetation.	identification by inference: lack of vegetation, fine regolith, light tone because of coarse lag or soil truncation.	light tone may be due to low soil moisture, secondary carbonates, low grass cover etc., also variation in reflection by lithology, angle etc. Small areas may be missed.	silt, clay and very fine sand.
Sh	surface	sheetwash	relatively smooth, sparsely or non vegetated slopes.	similar to above, often occurs with rainsplash.	similar to the above, but often with a connection to gullies.	similar to the above, usually occur interactively.	silt, clay and very fine sand
R	surface	rilling	similar to the above	U, V or ∨, colinear tracks less than about 30 cm depth, which are obliterated between runoff events. New rills developed with surface runoff.	inferred: often grade into gullies.	part of a continuum between sheet and gully erosion, defined by persistence (obliterated by subsequent storms), or depth (30 cm).	silt, clay and fine sand.
G	surface	gulling	sloping sparsely or non vegetated slopes, may headward erode into vegetated areas.	U, V, ◡ or ∨ shaped channels which are long relative to their width, semi permanent, usually greater than 30 cm depth.	linear erosion feature, easily ovservable once developed.	may be too small to be observed at the incipient stage. Small gullies may be obliterated by cultivation.	shape a function of grain size:
Ss	subsurface	lateral eluviation	any area with throughflow.	soil profile eluviation, infiltration of precipitation	no direct evidence, almost ubiquitous.	inferred.	colloids, dissolved solids
St	subsurface	tunnelling	fine regolith areas with throughflow.	subsidence, hydrolphytic vegetation along seepage lines, augering.	linear subsidence feature, may be darker tone because of soil moisture, may have hydrolphytic vegetation growth along seepage line.	do not necessarily have any surface expression, evidence may be obliterated by cultivation.	usually fine materlais (clay, sand and silt).
Cv	channel	incision (vertical erosion)	surface drainage network	stranding of flooplains, incision into bedrock and valley fill, scour holes.	inferred from flood plains with established vegetation.	vegetation may not be developed.	bed material, size inferred from channel character.
Cl	channel	lateral erosion	any area within reach of river flows, particularly at bends.	scroll bar development on the inside of meander bends with vegetation succession indicating rate of erosion, slumping banks, steep cut banks, trees with roots stranded on bars and along the banks.	the field recognition features may be visible.	vegetation, steep valley walls may obscure evidence in narrow channels, tack of vegetation may hinder inference.	bank material, size inferred from channel character.

MAPPING LEGEND: Process, severity of erosion, rate of delivery, type of material eroded, relative rate of contribution, type of material delivered to stream network.
Process: Sp = rainsplash, Sh = Sheetwash etc. above; Severity: L low, M medium, S severe; Delivery: L low, M moderate, H high; Material eroded: wash (solutes and fine clastics) mixed (wash and coarser clastics);
Relative rate: 1, low or negligible, 2, moderate, 3, major; Type: wash or mixed.
Example: ShWLL1W indicates sheetwash, wash load material, low erosion rate, low delivery rate and low or negligible contribution of wash load.

was used as the test area for this sediment supply analysis (Hudson, 1977). Excerpts from that analysis are now presented.

Air photo analysis suggested that the major fluvial sediment supply sources in the plains area of the Elbow River basin are from channel and riparian erosion. Lateral erosion of relatively high (15 to 25m) valley walls and river bank erosion are evident at numerous points throughout the plains reach (e.g., Figure 2.11) (Photo 2.9). Fine sand, silt and clay would be contributed from erosion of lacustrine valley walls and in part from the bank erosion. Bank erosion would be expected to yield predominantly gravel sized material. Infrequent erosion of valley wall till deposits would yield a large size range of material, from boulders to clays. Throughout the river reach there appear to be areas of aggradation (which are typified here by braided flow over the area of aggradation) and degradation (which are typified here by stranded, treed, floodplains). The channel is inferred to be in a quasi - equilibrium.

Upland erosion in the plains appeared to be limited to rilling and gullying of small areas along the tributary network and to deflation of cropped areas. The yield of sediment to the stream network appeared to be limited because the vegetation cover generally protects the highly erodible surficial material units, slopes are low, there are large areas of internal drainage and the ephemeral, vegetated channels appear to be inefficient (Photo 2.10).

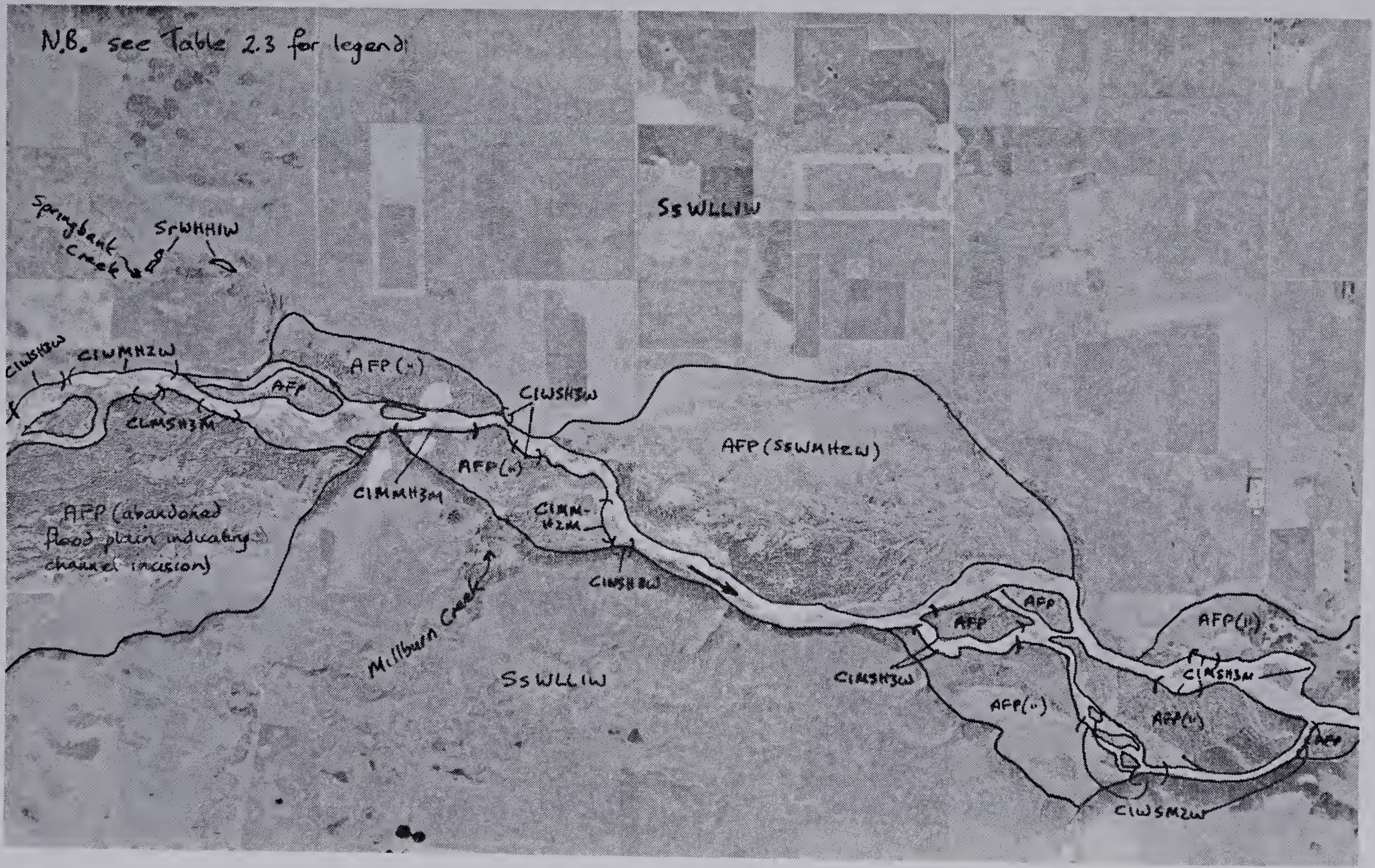

Figure 2.11 Sediment supply, plains area example

Photo 2.9 Erosion of glaciolacustrine and till valley walls occurs in the plains reach

Photo 2.10 Cullen Creek is a typical, grassed, ephemeral, plains area, tributary

The major erosion product from the uplands of the plains may possibly be material in solution from subsurface drainage.

In the foothills of the Elbow River basin air photo analysis suggested that the major fluvial sediment sources are channel and riparian erosion of floodplain, till and recessive bedrock exposures (sandstone and shale) (Photo 2.11). Lateral erosion appeared to be severe in bends with high till cliffs (Figure 2.12). Channel and riparian erosion appeared to be the major sediment source in the tributary streams as well (Photo 2.12). In addition, mass movements and surface wash processes may be important. However, many mass failures do not reach the channel network. Sheet and gully erosion is limited in extent, although the surficial material units are highly erodible, because of vegetation protection (Photo 2.13). What material is derived from upland and channel sources may be trapped in ponds associated with beaver dams and log jams before the sediment reaches the river (Photo 2.14). In some cases, such as in the McLean and Sylvester Creek area, high rates of erosion and efficient delivery may occur from well travelled seismic lines which run straight down slopes and across the stream network. Solute loads are inferred to be relatively large because, apart from a few denuded areas, the vegetation cover and thick regolith promote rapid infiltration and throughflow rather than surface runoff.

Air photo analysis suggested that the major sources of suspended sediment size materials in the mountains of the

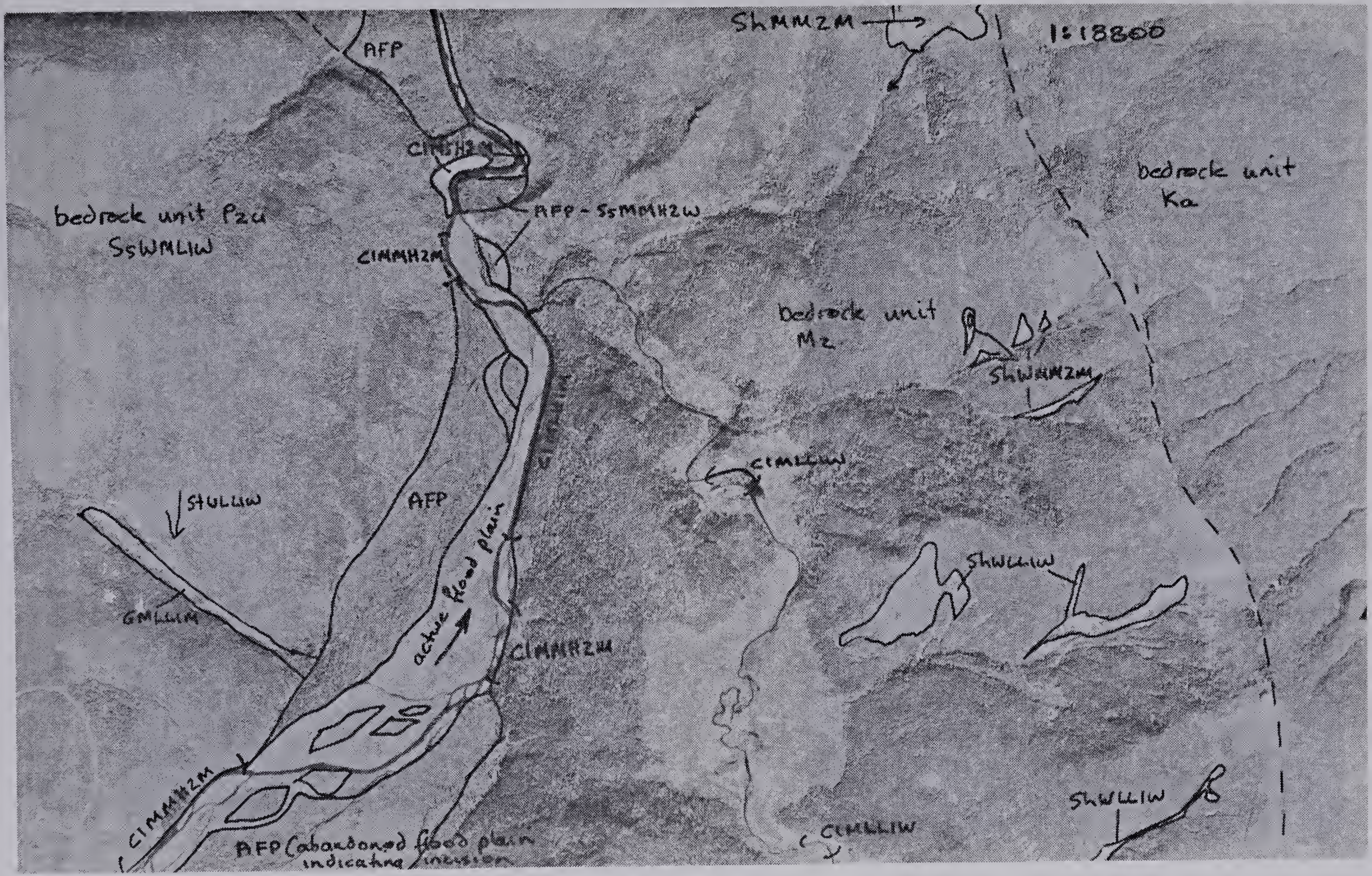

Figure 2.12 Sediment supply, foothills area example

Photo 2.11 Recessive bedrock exposures occur in the foothills

Photo 2.12 Lateral erosion of tributary streams may be an important sediment source

Photo 2.13 Vegetation tends to protect hillslopes in the foothills

Photo 2.14 Beaver dams often act as settling ponds in tributary streams in the foothills

Elbow River basin are from lateral river erosion of riparian till sections and from surface wash processes and mass movements of barren colluvial slopes (Photo 2.15). However, loads are probably relatively small because only a few sections of till - mantled hill slopes impinge upon the active river channel (Photo 2.16). Also, sediment delivery from upland sources is impeded by the tree cover on the lower mountain slopes (Figure 2.13). Erosion sites which do have direct links to the river are probably sediment supply - limited, because mechanical weathering of Paleozoic limestone and dolomite produces predominantly gravel sized particles, with miniscule amounts of finer material. The same factors which inhibit surface erosion processes, coarse textured and/or well vegetated regolith, also promote subsurface translation of water which should produce relatively high solute loads, because the regolith is calcareous.

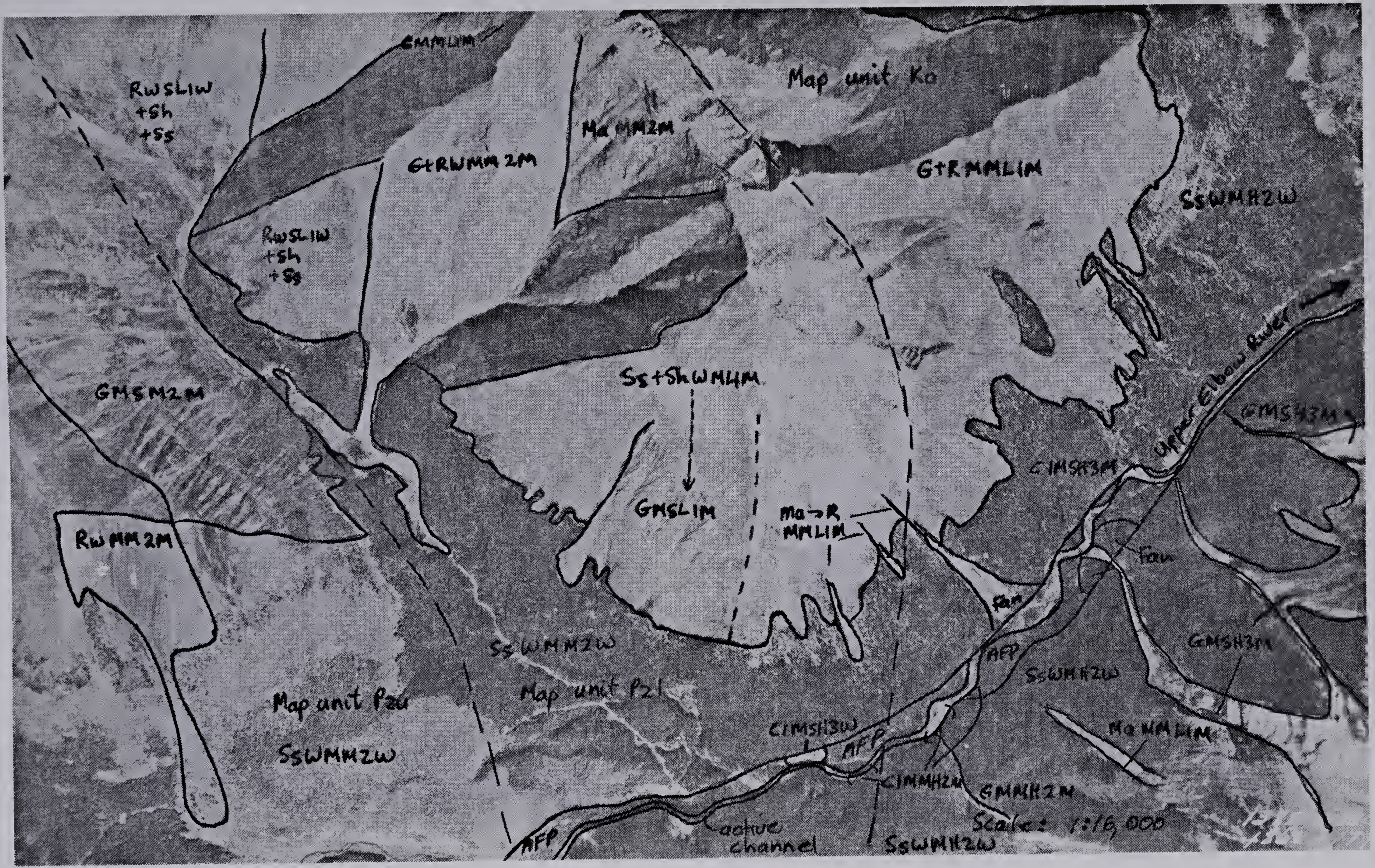

Figure 2.13 Sediment supply, mountain area example

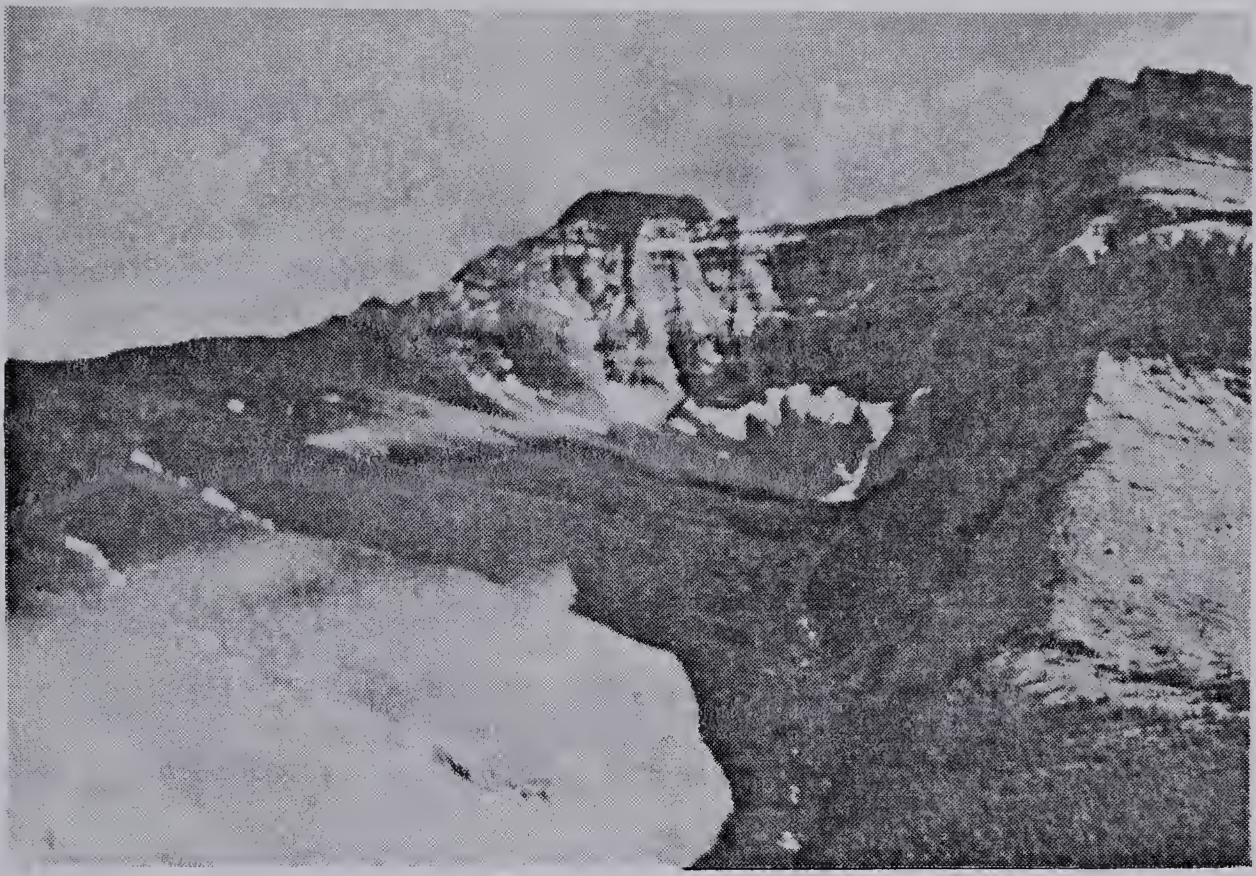

Photo 2.15 Tributary streams in the mountains originate in barren colluvial slopes

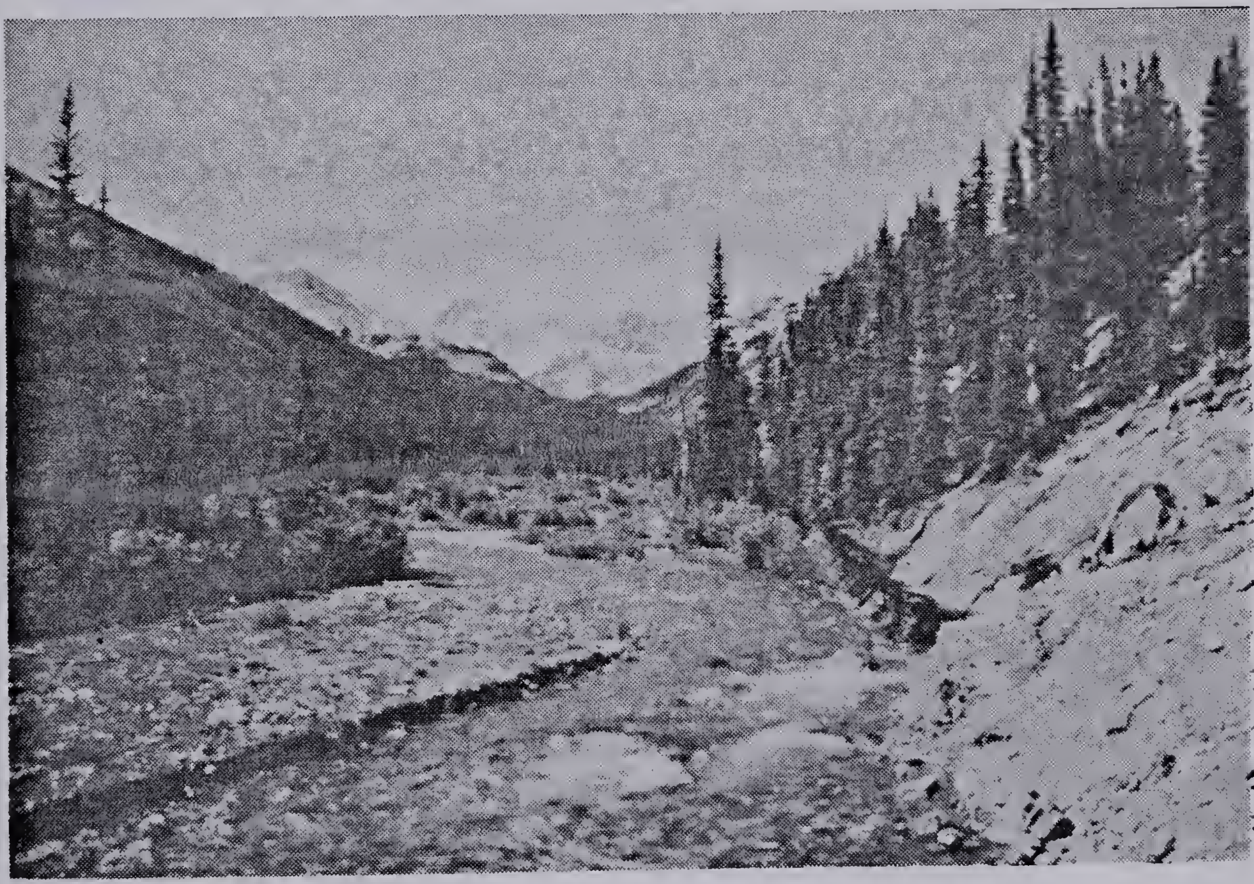

Photo 2.16 Erosion of till mantled valley wall deposits is infrequent in the mountains

3. HYDROLOGY OF THE ELBOW BASIN

A. STREAM FLOW DATA

There are four active recording hydrometric stations on the Elbow River River with a long period of record, three other recording stations with one to four years of record, and thirteen manual stage hydrometric stations with two years of record (Figure 3.1). The original gauging station, the Elbow River at Calgary (station 5BJ1), was renamed the Elbow River Below Glenmore Dam in 1932. This station has records since 1908. However, in the period 1934 to 1969 the discharges were estimated, by the City of Calgary, based on dam operations. Previous to, and after, that period discharges were estimated by Water Survey of Canada from stage records at a rated section.

Reservoir inflows are estimated for the Elbow River above Glenmore Dam (station 5BJ5), from City water use, changes in reservoir level, and an estimate of reservoir outflow. Since 1977 Water Survey of Canada have not published this data, because it is considered unreliable (M. Spitzer, pers. comm., 1978), at least during the peak discharge period.

The Elbow River at Bragg Creek (5BJ4), which is referred to as Bragg gauge to avoid confusion with Bragg Creek tributary, has high - flow stage records since 1923, mean daily gauge heights at a rated section from 1935 until 1948, and recording station data since 1948. Records are

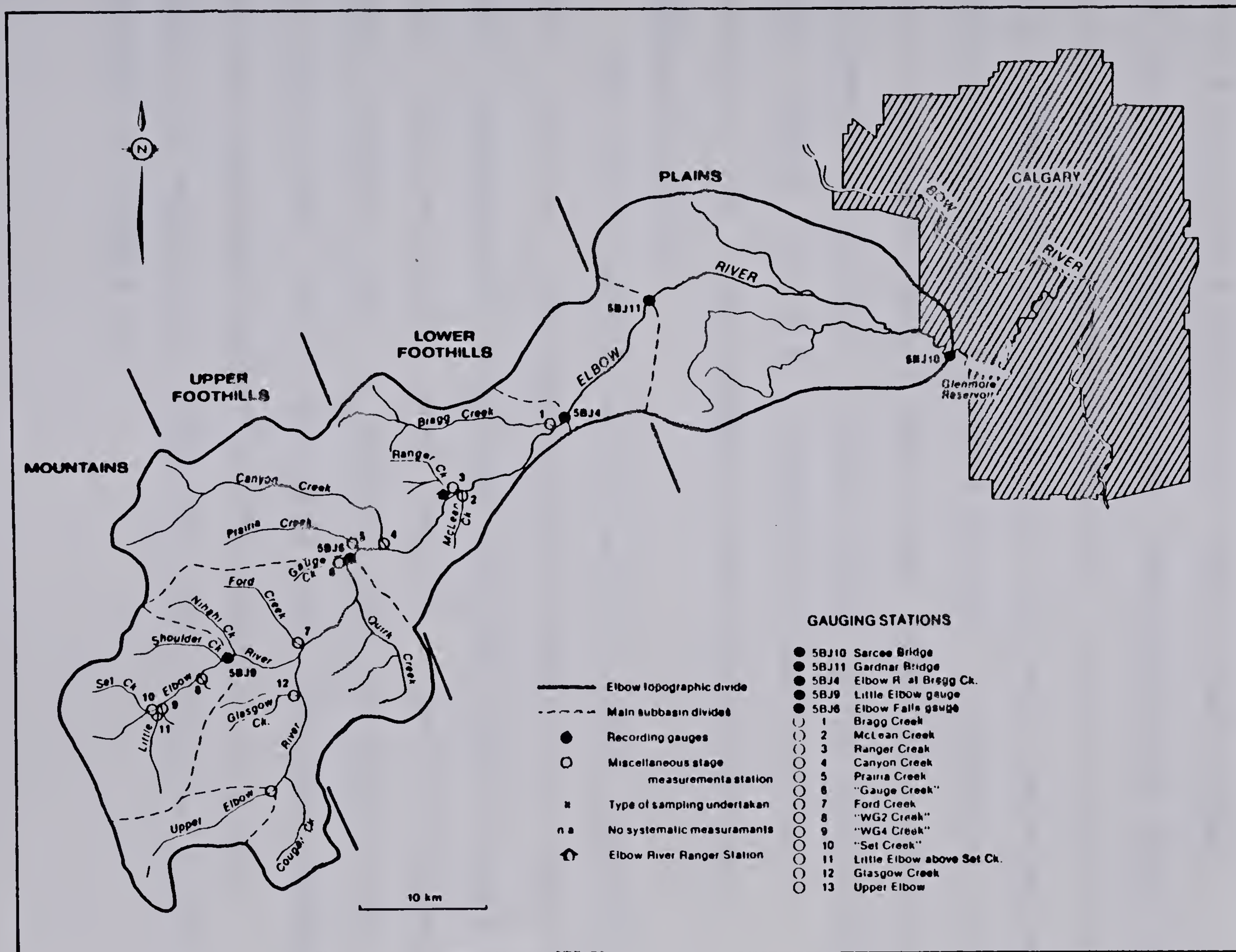

Figure 3.1 Elbow River basin hydrometric stations

usually obtained from March to October, inclusive. Since 1967 recording station data have been collected, on an annual basis, at the Elbow River above Elbow Falls (station 5BJ6), which is referred to here as Falls gauge for brevity. In 1978 the Little Elbow River above Nihahi Creek hydrometric station (5BJ9) was commissioned by Water Survey of Canada. This station is referred to as Little Elbow River gauge. Numerous other manual stations were installed and rated by the writer, for this project. In addition, in 1979, recording gauges were installed and operated by Alberta Environment at Gardner Bridge (5BJ11, km 48) and at Sarcee Bridge (5BJ10, km 19.5), immediately above Glenmore reservoir. The latter gauge has been maintained as a permanent station.

The runoff pattern at the manual gauging stations during snow melt was established by observation every few hours. Subsequently, stage was recorded three or more times a day. High water marks were used as a check of stage trends between observations. At McLean Creek and Ranger Creek continuous water level recorders were installed. However, they only operated intermittently because they were frequently vandalized. Thus, they are classified as manual stations in Figure 3.1.

B. BASIN RUNOFF

The long term, monthly, hydrometeorological records for the three long term, unregulated, gauging stations on the Elbow River illustrate the major features of the runoff regime (Figure 3.2). The three stations, Falls gauge (km 85, 435 km^2), Bragg gauge (km 60.4, 793 km^2), and the Elbow River above Glenmore Dam (km 11.8, 1210 km^2), sub - divide the Elbow River basin into three, almost equal, parts which represent the mountains and upper foothills, the mid foothills, and the plains, respectively (Figure 3.1). Specific features, alluded to in this section, are treated in detail in following sections.

During the winter months the river is ice covered and discharges are very low in relation to the remainder of the year (Table 3.1). Precipitation falls almost exclusively as snow and air temperatures are, on average, well below freezing (Figure 3.2). The storage of precipitation as snow, and the rapid release of the snow during melt, effectively redistributes several months of precipitation into the short snow melt period in the foothills and mountains. In the plains very warm temperatures, associated with chinooks, may remove the snow cover on several occasions in any year prior to the spring melt. Chinooks may also occur in the foothills and mountains of the Elbow River basin (Buckler, 1968). Snow melt runoff occurs at progressively greater elevations within the basin over a period of days to weeks. Thus, winter in effect persists in the upper basin until May or

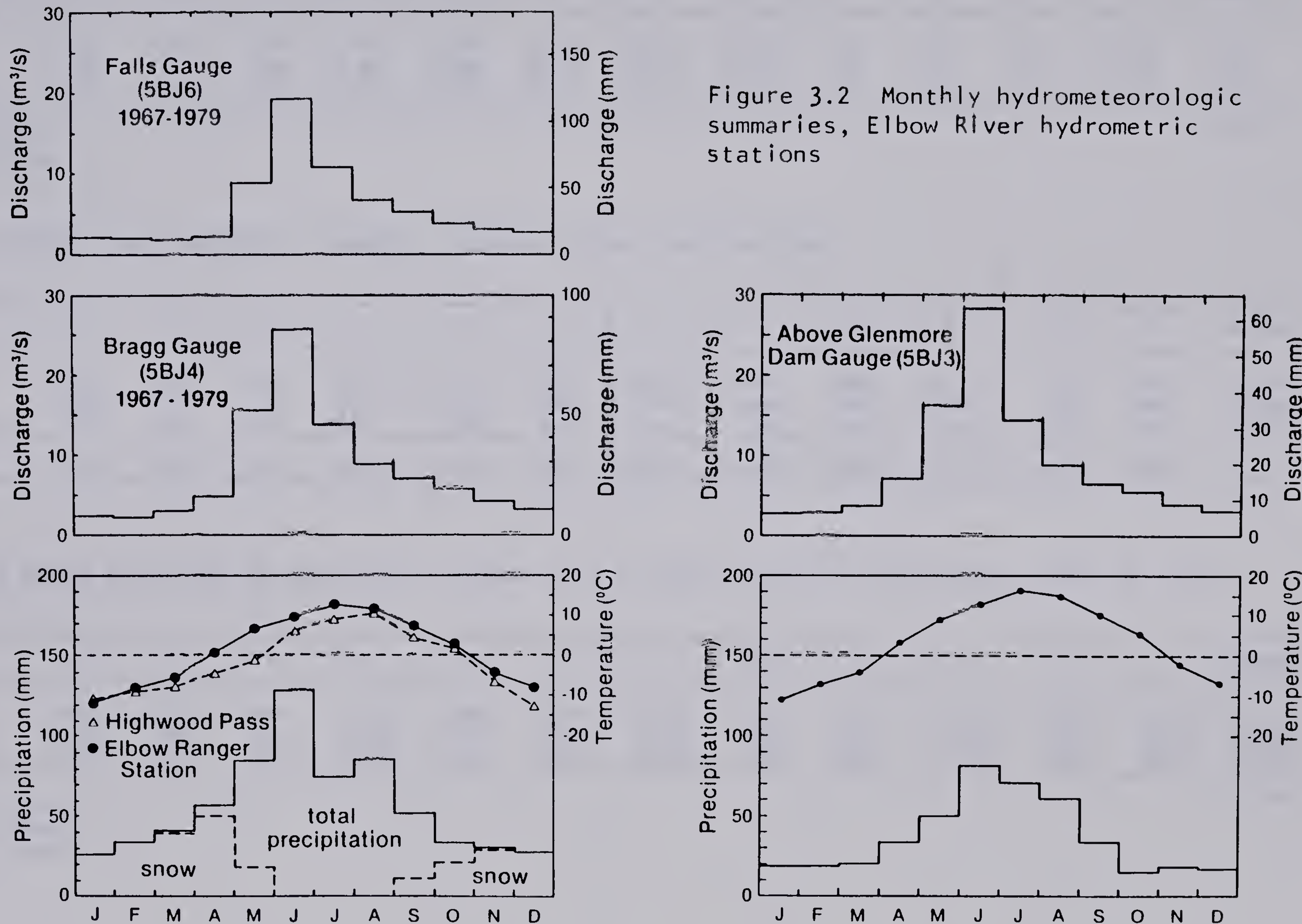

Figure 3.2 Monthly hydrometeorologic summaries, Elbow River hydrometric stations

Table 3.1 Mean monthly runoff, Elbow River hydrometric stations

STATION	J	F	M	A	M	J	J	A	S	O	N	D	ANNUAL
Elbow Falls	2.13	1.98	1.86	2.17	8.81	19.14	10.96	6.76	5.16	3.95	3.13	2.53	5.72
Bragg Creek	2.46[1]	2.45[1]	3.12	4.80	15.62	25.72	13.72	8.96	7.07	5.72	4.13[2]	3.17[2]	8.08
above Glenmore	2.76	2.97	3.71	7.03	16.16	28.29	14.42	8.92	6.50	5.50	3.90	3.02	8.60

[1] 3 years of data
[2] 4 years of data

(m^3/s)

Table 3.2 Mean net monthly runoff, Elbow River sub basins

STATION	J	F	M	A	M	J	J	A	S	O	N	D	ANNUAL
Falls guage	5671[1]	4821[1]	5071	5648	24173	51766	30140	18767	13685	10738	8109[1]	6787[1]	185376
Bragg Creek	922[2]	1171[2]	3172	6764	18771	18421	7478	5867	4881	4803	2622[3]	1687[3]	76559
above Glenmore	794	1261	1703	5811	333	3142	1006	-748	-1715	-803	-624	-386	14050
TOTAL	7387	7253	9946	18223	43277	73329	38624	23886	16851	14738	10102	8088	da m^3

Table 3.3 Mean monthly discharge, Elbow River hydrometric stations, 1967 to 1979

STATION	J	F	M	A	M	J	J	A	S	O	N	D	ANNUAL
Elbow Falls	5671[1]	4821[1]	4996	5625	23588	49575	29343	18110	13380	10595	8109[1]	6789[1]	180602
Bragg Creek	6593[2]	5992[2]	8344	12438	41851	66652	36748	24001	18312	15322	10731[3]	8476[3]	255460
above Glenmore	7387	7253	9946	18223	43277	73329	38624	23886	16851	14738	10102	8090	271706

[1] 9 years of record
[2] 3 years of record
[3] 4 years of record

da m^3

June, while further downstream monthly discharges increase in March and April (Figure 3.2). As a result, proportionately more of the runoff is generated from the lower foothills and plains area in early spring.

Approximately 60% of the total annual discharge of the Elbow River basin occurs in May, June and July. The spring peak discharge of the whole basin occurs in May or June. On average, 56% of the May runoff comes from above Falls gauge, 43% is derived from the Bragg gauge sub - basin and less than 1% is derived from the lower basin between Bragg and Glenmore gauges (Table 3.2). The small proportion derived from the lower basin is, in part, attributed to the Calgary City Water Works Department underestimating the flood peak discharges for the Elbow River above Glenmore Dam (Appendix 2). In June, the flood generation zone shifts further up - basin and the Falls gauge sub - basin produces 71% of the monthly discharge, the Bragg gauge sub - basin produces 25%, and the Glenmore sub - basin produces about 4% of the monthly discharge.

Although the proportion of the monthly discharge derived from the upper basin above Falls gauge remains at about 70% of the monthly discharge at Bragg gauge (Figure 3.3), the area where the flood peak is largely generated, changes. Small flood peaks tend to be generated from the mountains and upper foothills and larger flood peaks include substantial runoff from the mid and lower foothills and plains. That is, the relative importance of the flood

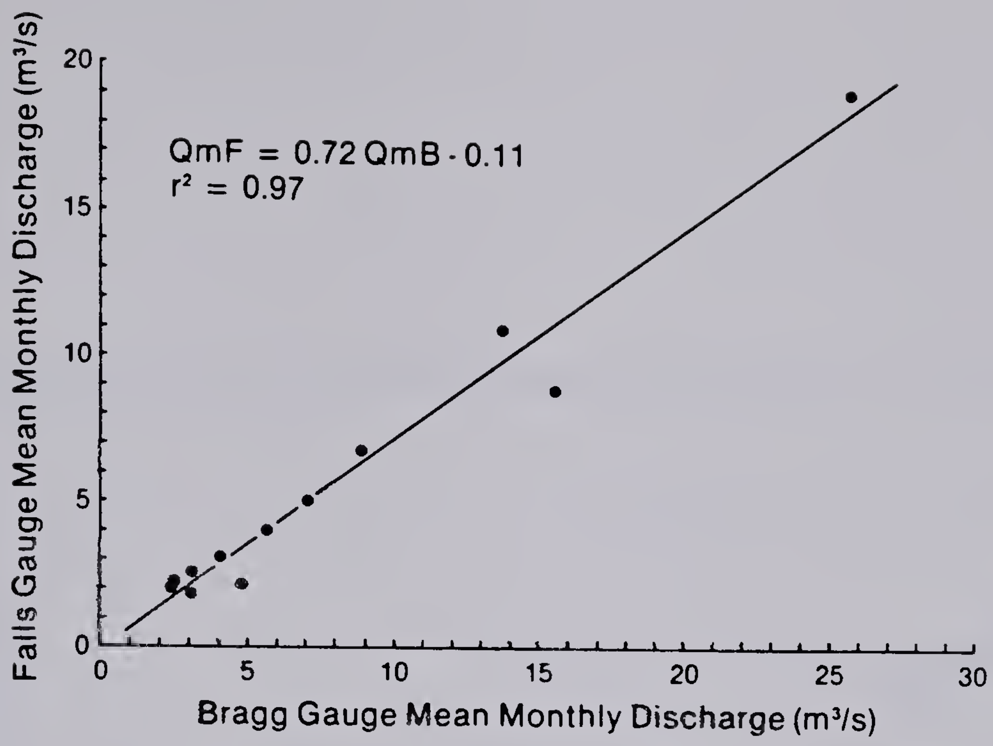

Figure 3.3 Mean monthly discharge at Falls gauge against Bragg gauge, 1967 - 1979

generation zone shifts eastwards to involve more of the basin as the flood magnitude increases (Figures 3.4 and 3.5). These points are developed later in this chapter and in Appendix 2.

Summer recession usually begins in early June, and discharge decreases rapidly to reach a low in late winter. Several, generally smaller, rainfall peaks can occur during the snow melt recession period, which persists through July. Summer rainfall produced the annual maximum discharge in 7 years from a 45 year record at Bragg gauge. These summer floods are generally relatively small, with a 2.0 year return period, on average, with a maximum of a 4.4 year return period, in the annual series.

During summer, and through winter, discharge increases downstream from the mountains and the foothills. However, a

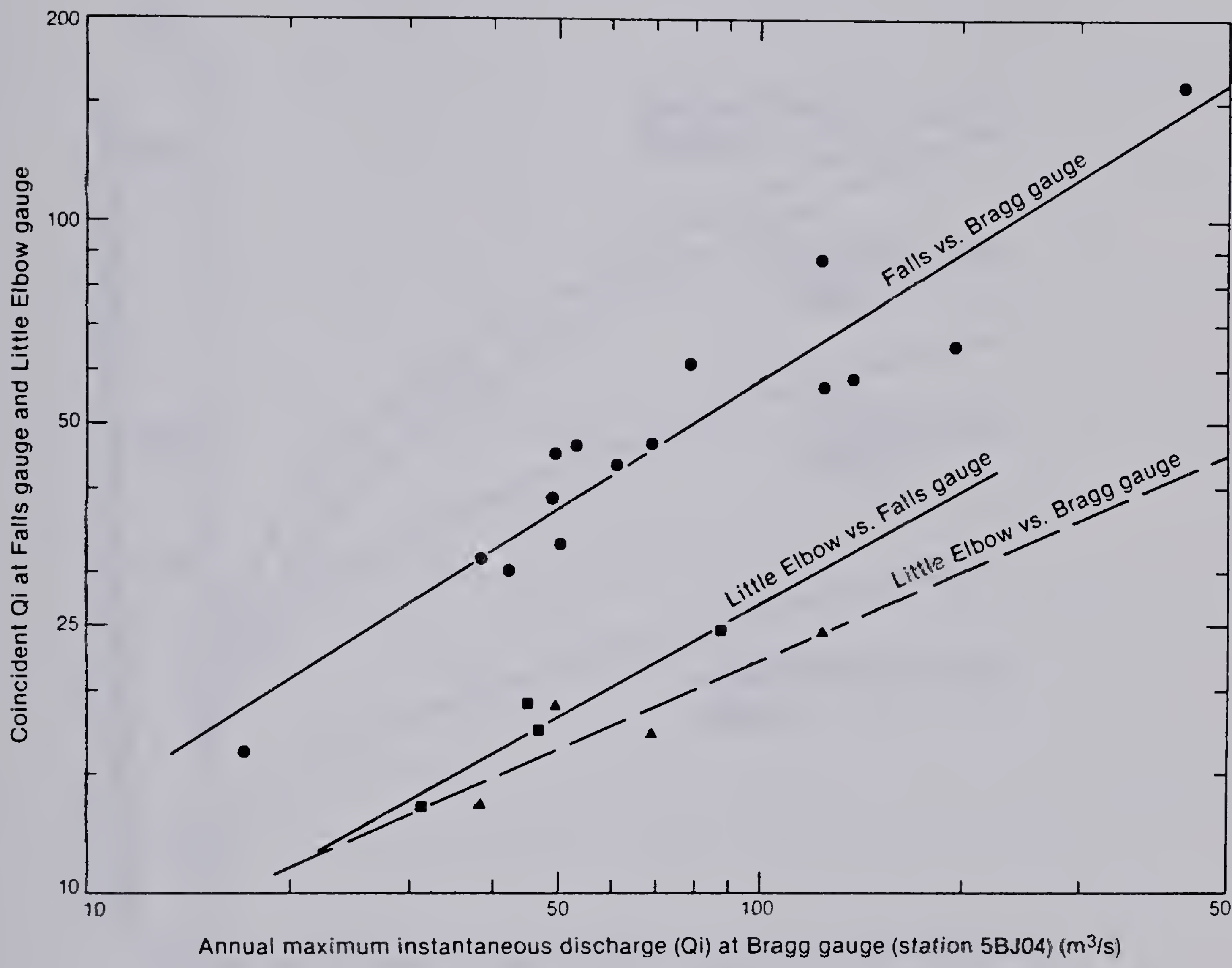

Figure 3.4 Falls gauge and Little Elbow gauge annual maximum instantaneous discharge (Qi) against Bragg gauge Qi

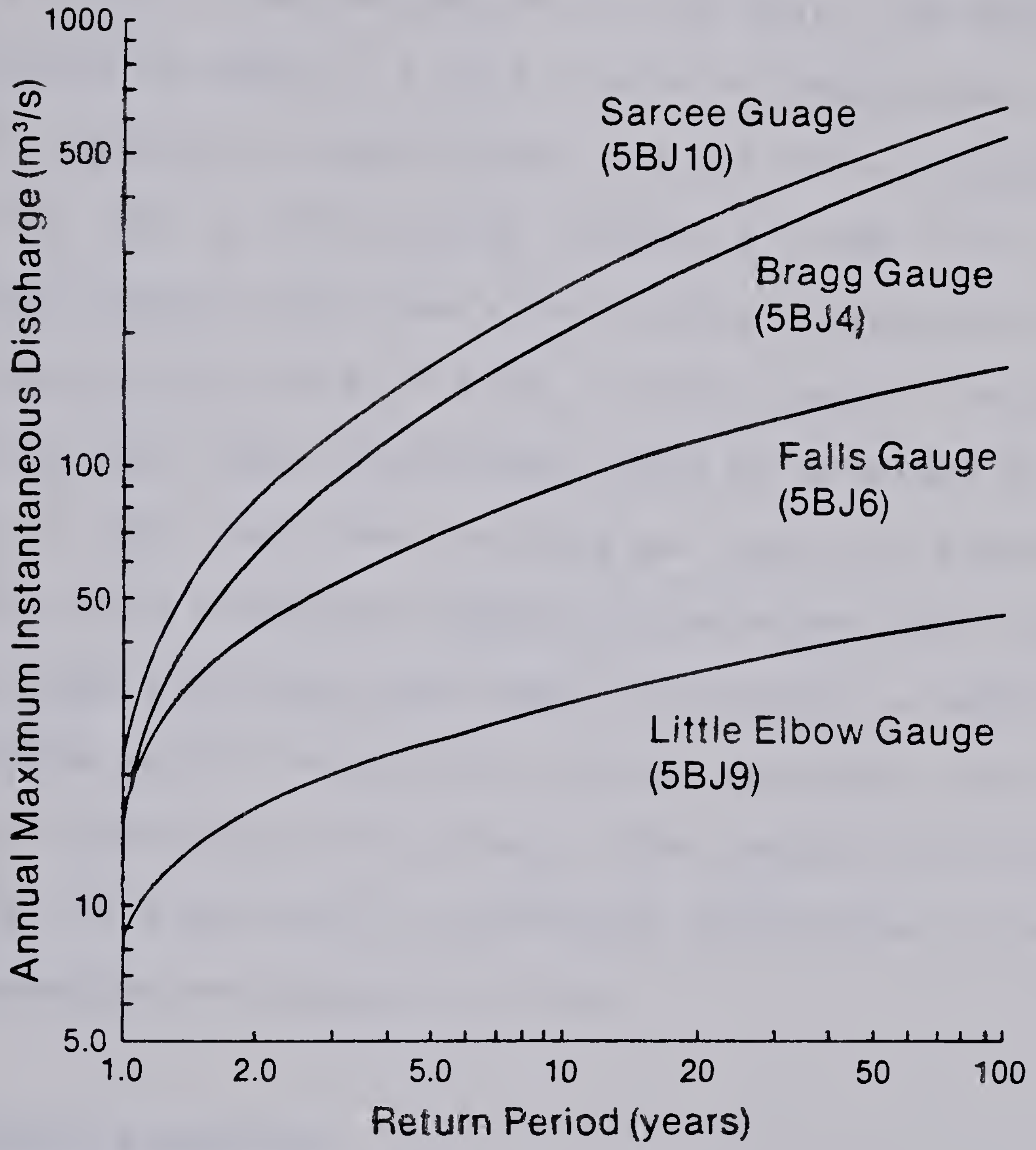

Figure 3.5 Flood frequency estimates, Elbow River basin hydrometric stations

major decrease in discharge occurs through the plains reach between Bragg gauge and Glenmore Reservoir during the summer and early winter (Figure 3.1; Table 3.2).

A flow duration analysis shows that discharge is low for all but a limited period of the year. The median discharge is about 7.1 to 8.2 m^3/s at Bragg gauge, depending on the period of record used, and 4.4 m^3/s at Falls gauge for the 1967 to 1979 period (Figure 3.6 and 3.7). The average annual flood has a mean daily discharge of 74.7 m^3/s at Bragg gauge and 42.4 m^3/s at Falls gauge, for the 1967 to 1979 period. These discharges would be exceeded 0.5% of the time, or for less than two days per year, on average. A plot of the daily discharges against time shows that, typically, discharges are relatively small in winter, a rapid rise in discharge occurs in spring; and the recession from the peak annual discharge is very rapid. The regime is illustrated in Figure 3.8 where daily discharges, precipitation and temperatures are shown for 1978.

C. WINTER HYDROLOGY

Winter runoff records are available for the lower basin for the period 1935 to 1977 (station 5BJ5). However, coincident winter records at Bragg gauge (5BJ4), are available for only January and February 1972, 1978 and 1979, and for November and December of 1971, 1977, 1978 and 1979.

Given the limitations of the winter data, two trends are suggested. First, in early winter (November and

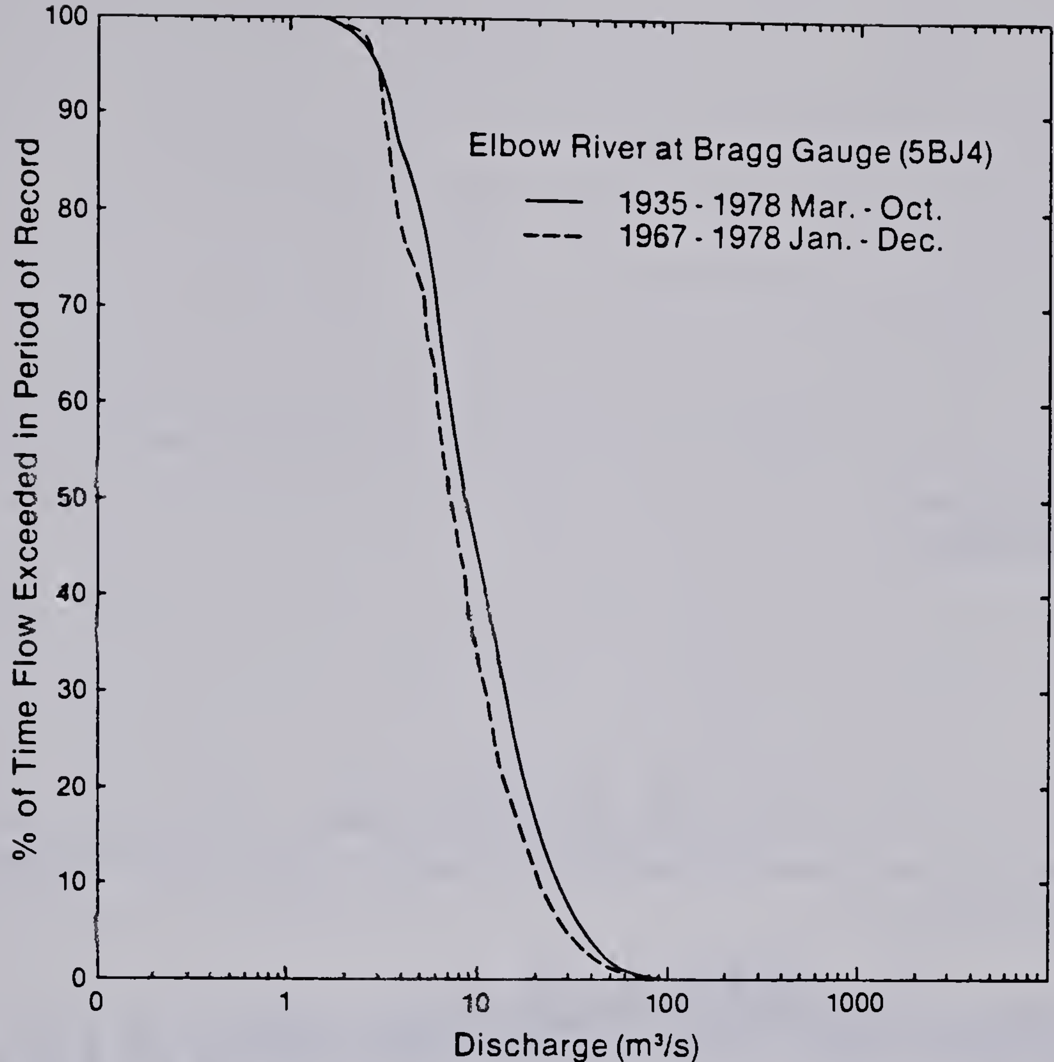

Figure 3.6 Flow duration curve, Elbow River at Bragg gauge

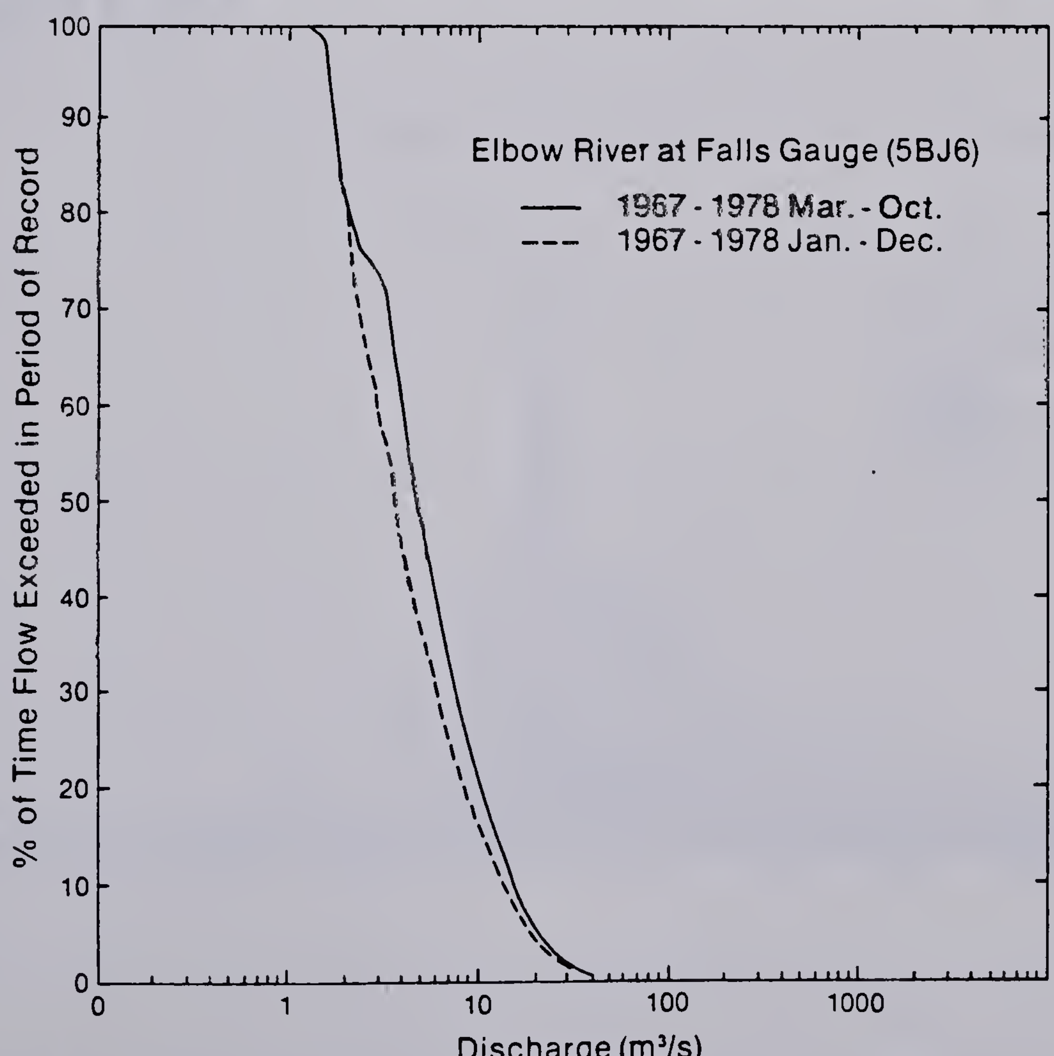

Figure 3.7 Flow duration curve, Elbow River at Falls gauge

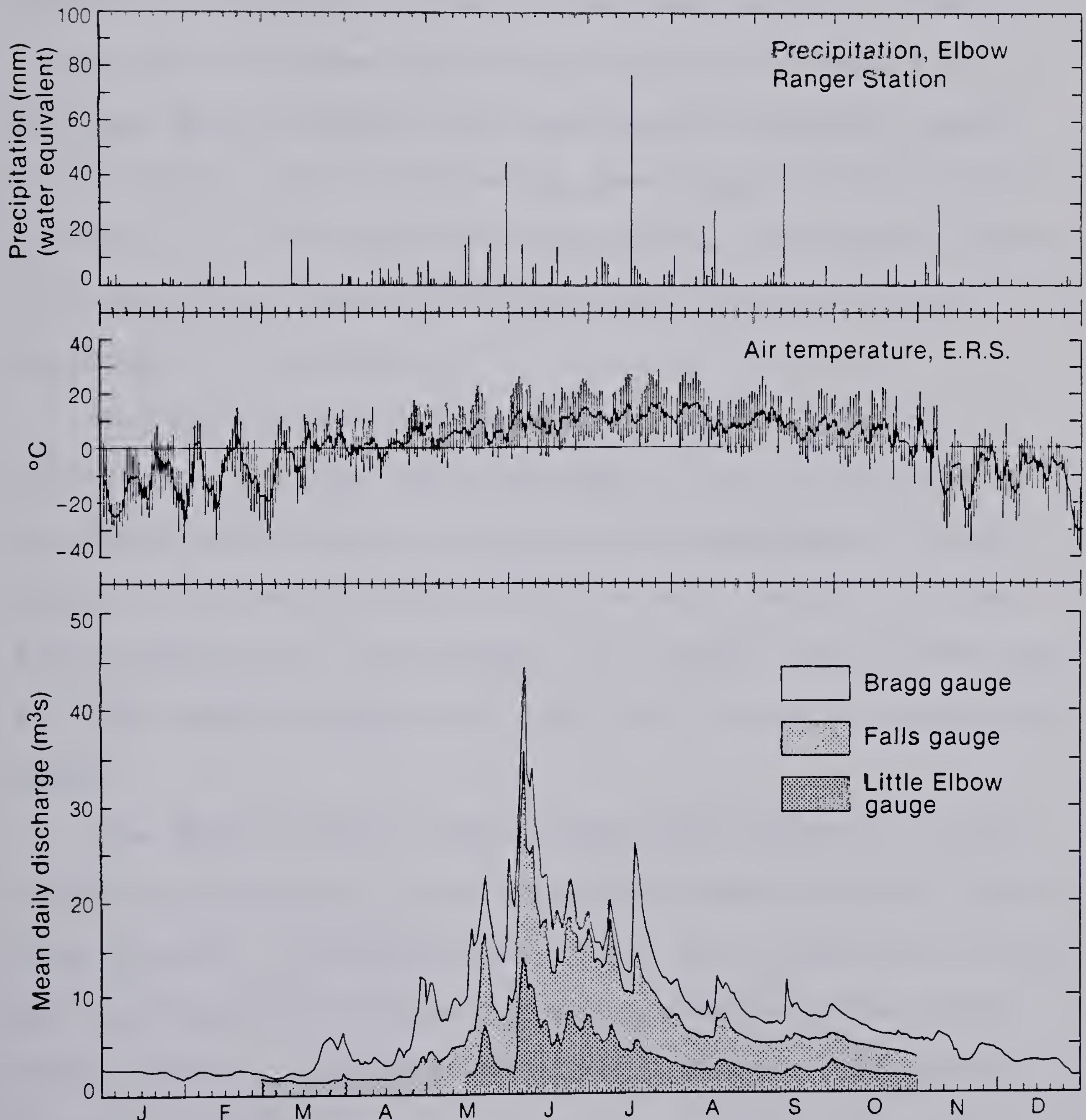

Figure 3.8 Elbow River at Bragg gauge hydrometeorology, 1978

December) the lower basin is in a major water - deficit situation (Laycock, 1957). Hence snow melt is stored and the decrease in discharge downstream (Tables 3.3 and 3.4) of about 5.5% of the mean daily discharge (Table 3.1), may be due to losses to the atmosphere, bank recharge or deep seepage, or perhaps a discharge estimate error. In late winter the discharge recorded at Glenmore increases while the mean daily discharge at Bragg gauge decreases (Table 3.1). The net gain in discharge downstream is attributed to snow melt in late winter during numerous warm spells (Table 2.4) even though much of the snow melt is lost to the atmosphere or recharge.[1]

During the period November to February, limited coincident flow data shows that 65% to 79% of the flow of the Elbow River system is derived from upstream of Falls Gauge at an average rate of 2.44 m^3/sec (Table 3.1). The Little Elbow River contributes, on average, 31% of the flow at Falls Gauge (Table 3.4), from 29% of the area above Falls gauge.

The daily flows at Falls Gauge are typically low and steady during winter, with a tendency toward gradual decline from November to early spring (Table 3.1). Responses to warm and particularly cold spells, as recorded at Elbow River Ranger Station, appear to be minimal (Figure 3.8). This

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[1]Average January and February precipitation = 37.6 mm (Table 2.5); specific drainage area = 388 km^2, thus precipitation input = 14588 da m^3. Average runoff from the lower basin for this period = 2055 da m^3. Thus, the runoff coefficient = 2055/14588 = 0.14.

Table 3.4 Percentage of total annual discharge occurrring in each month, Elbow River hydrometric stations

STATION	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
Little Elbow[1]	3.04	2.63	3.24	3.90	12.74	25.52	16.27	9.85	7.91	6.98	4.24	3.59
Elbow Falls[2]	3.14	2.67	2.77	3.11	13.06	27.45	16.25	10.03	7.41	5.87	4.49	3.76
Bragg Creek[3]	2.58	2.35	3.27	4.87	16.38	26.09	14.39	9.40	7.17	6.00	4.20	3.32
Glenmore[4]	2.72	2.67	3.66	6.71	15.93	26.99	14.22	8.79	6.20	5.42	3.72	2.98

[1] two years of record 1978 & 1979; winter discharge estimated from Falls gauge data
[2] nine years of record for winter months
[3] three to four years of record for winter months
[4] 1967 to 1977 inclusive

Table 3.5 Estimated mean daily discharge, selected Upper Elbow River basin tributaries, spring and summer, 1979

Station	May	June	July	August	Area km^2
Falls Gauge	7.99	10.70	6.58	4.32	307
Little Elbow Gauge	2.87	4.71	2.89	1.91	128
Upper Little Elbow	0.57	0.9	0.4	0.25	16.3
WG 2	0.07[1]	0.06	dry		3.9
WG 4	0.20[1]	0.18	0.01	dry	10.9
Set Creek	1.6	2.7	1.9	1.2	53.2
Upper Elbow	1.72	2.83	1.73	1.15	65.3
Cougar		1.1	.7	0.3	29.6

Discharge estimated (m^3/s) based on miscellaneous streamflow measurements and published records (Water Survey of Canada, 1980).

[1] frozen until the last week of May, mean daily discharge given for the last week of May.

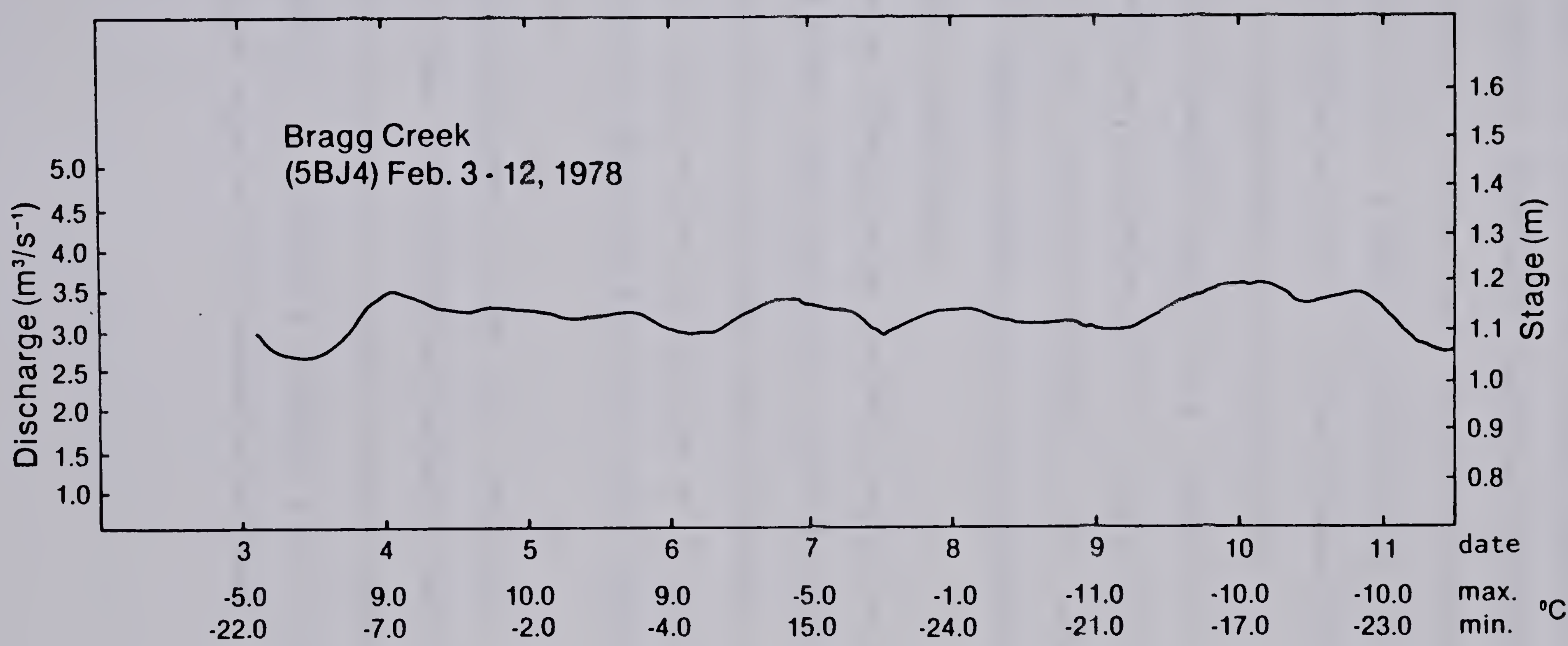
limited response may be due to the protection of the snow pack by low temperatures in the mountains and shading in the valleys.

Winter flows at Bragg gauge are invariably reported to be low and steady with a monotonic decline from the late summer rains through winter, with the lowest flows occurring in February. However, the 1978 and 1979 hydrographic traces show that considerable fluctuations in stage occur during this period.

The recorded stages and reported mean daily discharges have a less than ideal relationship. The reported discharges are often at variance with the trends and magnitudes of the recorded stage and bear no parallel with temperature fluctuations. A winter rating curve was developed by the writer. Published mean daily discharges were plotted against mean daily water levels and a curve was fitted through the data, after obvious outliers were excluded. When this rating curve is applied to the record of river stage, a reasonable relationship between discharge and meteorological conditions is evident (Figure 3.8). River ice conditions usually prevail from mid November until early April.

From the pattern of the hydrographic traces it is evident that the considerable variation in recorded stage may be attributable to two phenomena: (a) a variation in cross - sectional area of the stream as freeze - up and ice removal takes place, and (b) hydrologic responses to meteorological events (Figure 3.9).

Figure 3.9 Elbow River at Bragg gauge hydrograph, February 1978



Generally the winter hydrographs have a more gradual response to meteorologic controls than do the spring or summer hydrographs. Lows in discharge tend to correspond with extremely low temperature periods. Higher flows appear to have two components: (a) a gradual rise and fall in discharge over several days with (b) a superimposed mid afternoon rise which corresponds with maximum daily temperatures exceeding freezing point (e.g. February 4 - 8, 1978; Figure 3.9). This may represent melt with a division into stream flow and into temporary storage in the snow pack (Snow Hydrology, 1977) which results in a protracted but attenuated hydrographic response.

Winter discharge at Bragg gauge is derived primarily from the upper Elbow River basin and not the mid foothills zone. For the period 1967 to 1979 data from different years show that the average total winter discharge at Bragg gauge was 31,792 da m^3, of which 80% was derived from upstream of Falls gauge (Table 3.3).

The relative contributions of the sub - basins, however, change with time. Limited data suggest that the mid foothills have relatively greater flows than the upper basin in early winter (Table 3.1), probably because the tributaries do not freeze up as soon as the mountain streams.

D. SPRING RUNOFF

Spring runoff in the Elbow River basin is divisible into two distinct periods: an early spring period of relatively low discharges, where melt progresses up basin, and the relative contribution to the system discharge from each physiographic zone changes; and, the main spring flood, which is dominated by runoff from the middle and upper basin.

When each sub - basin is isolated, it is apparent that the bulk of the sub - basin total annual discharge occurs in April in the plains, in May in the lower foothills, and in June in the upper foothills and mountains (Figure 3.10). Throughout these periods the bulk of the discharge is still derived from the foothills and mountains, but the proportion derived from these areas is less than at any other time (Table 3.1).

Laycock (1957: 33) suggests that "The marginal and variable flow of the lower basin is received largely from snow melt waters and above normal spring rains in April and May." This situation appears to be the case over the long term (Table 3.2) and snow melt is thought to be the source of the excess runoff from the Gardner sub - basin in April 1979, because of the correspondence between discharge and meteorological conditions (Figure 3.11) and because the groundwater input is estimated to be minimal during this period (Appendix 3).

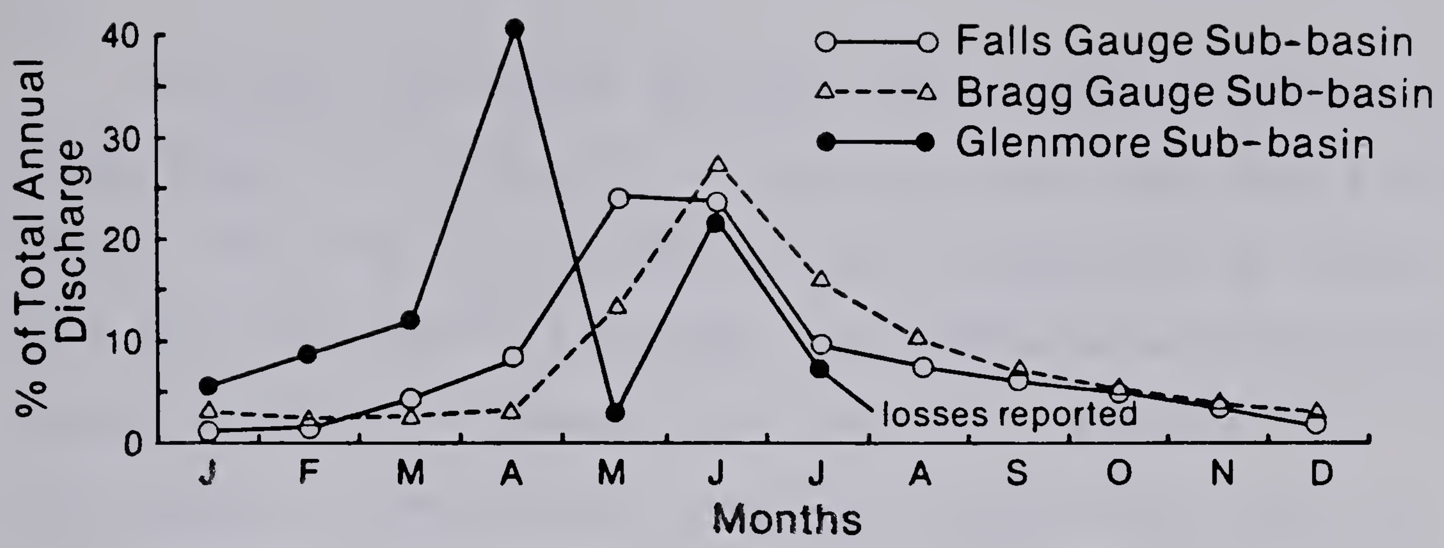


Figure 3.10 Mean monthly proportion of the total annual discharge, Elbow River sub-basins, 1967 to 1979

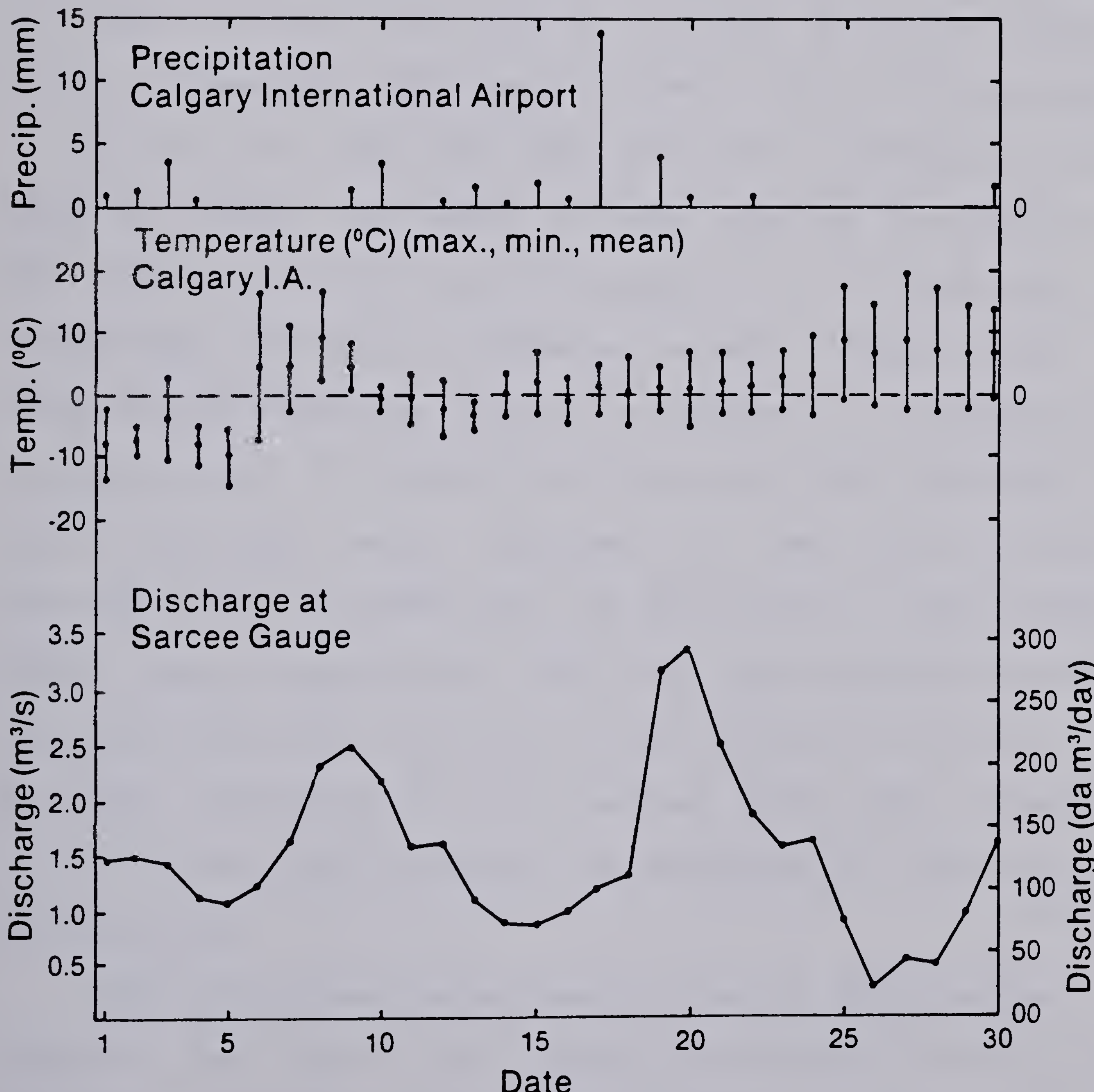


Figure III.11 Net runoff, Lower Elbow river basin, April 1979

Although the winter and main spring flows reported at Bragg gauge are primarily derived from the basin above Falls Gauge, the early spring hydrograph is dominated by runoff from the mid foothills because the discharge at Falls gauge is increasing at a lesser rate (Table 3.3 and 3.4). An examination of mean daily discharge figures from 1967 to 1979 at Bragg gauge and Falls Gauge shows that the first "major" rises in recorded mean daily discharge occur from a few days to a few weeks earlier at Bragg gauge than Falls Gauge. To quantify this lag the lowest flow at Bragg gauge in each year was noted and the concurrent discharge at Falls Gauge was noted. The number of days required to achieve a 50% increase in flow from the respective low flows were established from daily discharge records. "Significant" early spring discharge occurs, on average, 17.7 days earlier at Bragg gauge. To double the discharge (100% increase) the lag is 18.5 days, which indicates: (a) that the main flow from the mountains dominates the discharge at Bragg gauge and (b) that "significant" flow from the mountains is due to the main snow melt flood which rises sharply from the preceding relatively low flow period. There does not appear to be a relationship between the magnitude of the flow and the lag time.

The spring peak instantaneous flow at Bragg gauge occurs in late May or June. Usually the spring flood peak is the maximum instantaneous flow recorded for the year (38 out of 45 years of record and 64 out of 74 years at the Elbow

River below Glenmore Dam). About one quarter of the total annual discharge (t.a.d.) at Bragg gauge occurs in June and almost 60% of the t.a.d. occurs during the snow melt period May, June and July (Table 3.4). The hydrographic trace is characterized by a distinct diurnal rise and fall associated with snow melt during this period. The pattern may be disrupted by rainfall. The main hydrographic rise tends to occur at the same time and to effect all of the stations to some degree. The small mountain tributaries flow for a limited period (in the order of six weeks) (Table 3.5), with the bulk of the runoff occurring in a rapidly rising, markedly diurnal, snow melt runoff flood which produce relatively high flows (Figure 3.12 and 3.13) for only a few days. The larger mountain tributaries and foothills streams, with the exception of Canyon Creek, are perennial and exhibit a similar distribution of runoff to the Bragg gauge regime.

The relative importance of flood generation areas appears to change with the magnitude of the flood event. Small floods tend to be generated predominantly from the mountains and upper foothills. As the magnitude of the flood increases, the lower foothills and plains generate additional runoff. Large floods are generated when substantial frontal rainfall occurs at the time of spring snow melt. Rainfall is at a maximum along the eastern margin of the foothills and tends to decrease both eastwards and westwards. Thus, the relative importance of the upper basin

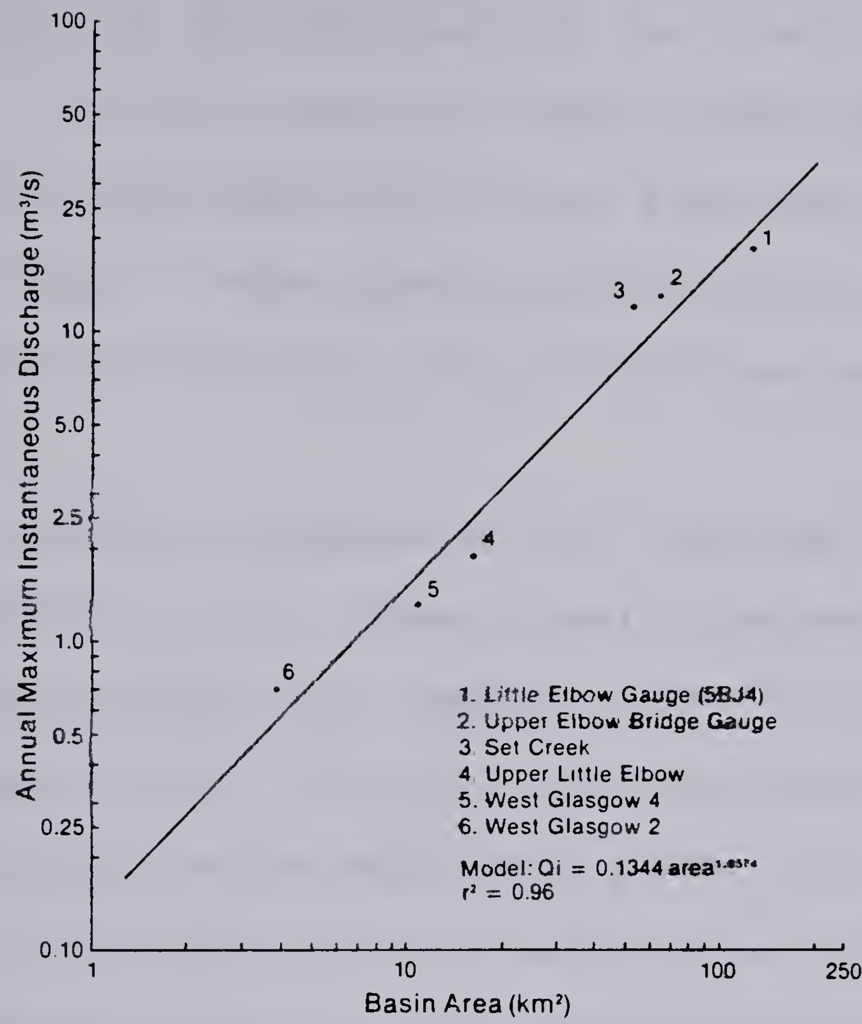


Figure 3.12 Upper Elbow River basin tributaries, annual maximum instantaneous discharge against drainage area, 1979

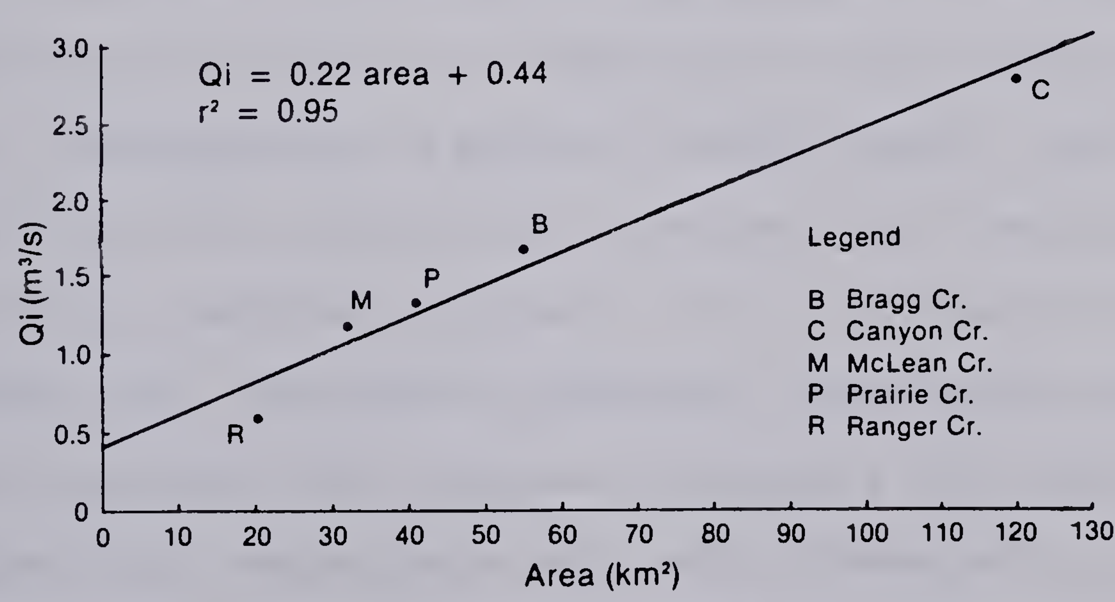


Figure 3.13 Foothills tributaries, annual maximum instantaneous discharge against drainage area

declines as the magnitude of the flood increases (Figure 3.4). Generally warm air temperatures rapidly melt the substantial snow pack over large areas and rainfall runoff is also basin - wide. These conditions, if optimal, are expected to generate the probable maximum flood (Buckler, 1968).

Varying scenarios of the "optimum flood producing" condition prevail. Usually melt progresses up - basin so that the Bragg sub - basin peak occurs several days before the main runoff from the mountains. Rainfall in the foothills may produce additional peaks. Low flood magnitude years are typified by a limited available snow pack and a protracted period of melt, due to a slow warming trend, rather than rapidly rising temperatures after a cold winter and cool early spring.

There may be a complex response to spring precipitation. For example, in 1978 the mountains were frozen - in until early May. Increasing temperatures in mid May produced significant increases in discharge at all stations (Figure 3.8). Precipitation resulted in major runoff from the mid - foothills and a considerably smaller rise from the upper foothills. The mountains had little or perhaps no response (Figure 3.5). Increased cloudiness, associated with precipitation which fell as snow, produced a decrease in mountain air temperatures and, as a consequence, a decrease in snow melt. As the temperatures rose over the following few days melt of the new snow in the upper basin produced a

significant hydrographic rise.

E. SUMMER RUNOFF

The post peak spring runoff period is characterized by the dominance of flows from the upper basin. Discharge decreases dramatically from the peak flow period and the correspondence between air temperature and runoff is complicated by major hydrograph rises which are a result of rainfall (Figure 3.8). The rainfall peaks are superimposed on a receding flow hydrograph, which has a small diurnal fluctuation, which decreases in amplitude into the summer as the alpine snow melt zone recedes to progressively greater elevations.

Two types of rainfall event seem to be important: local convectional showers, which tend to be generated over the foothills and migrate onto the plains; and, frontal rainstorms, which tend to have their center of maximum precipitation in a long narrow strip along the edge of the foothills, with decreasing precipitation both eastwards and westwards (Appendix 2). Convectional showers are usually very local in extent, and are frequently not recorded by the sparse meteorologic network.

Local convectional showers have a variable response across the basin. The ephemeral mountain and plains tributaries did not respond to the short duration, intense rainfalls observed in 1978 and 1979. The response of the larger perennial tributaries in the mountains appears to be

limited to an in - channel precipitation response.

The perennial foothill streams all show some response to short duration rainfall, but the response is quite varied. Ranger Creek, for example, had little short term response to these storms while nearby McLean, Sylvester and Gauge creeks were very "flashy" (Figure 3.14). The response is explicable in terms of basin character. Ranger Creek is largely spring fed, well vegetated and has extensive ponded areas behind numerous beaver dams. McLean, Sylvester and Gauge creeks do not have beaver dams and they produce significant surface runoff, during intense rainstorms, from numerous well travelled roads and trails which cross the streams.

Frontal rainstorms may produce significant hydrographic rises. In the 45 year period 1935 to 1979, seven annual maxumum instantaneous discharges occurred in July or August. Numerous smaller peaks also occur (Figure 3.8). The largest of these flows (1951) had a 4.4 year return period and the average return period of rainfall peak annual instantaneous flows is 2.0 years.

A twelve day period in July, 1978, illustrates the hydrologic response of the upper basin and foothills to a major frontal rainstorm, which occurred almost simultaneously over the whole basin. Rainfall was maximum in the mid foothills (Elbow River Ranger Station, 95.20 mm) and decreased eastward, westward, and with elevation (Glenmore Dam, 1067 m, 25.60 mm; Kananaskis, 1390 m, 49.60 mm;

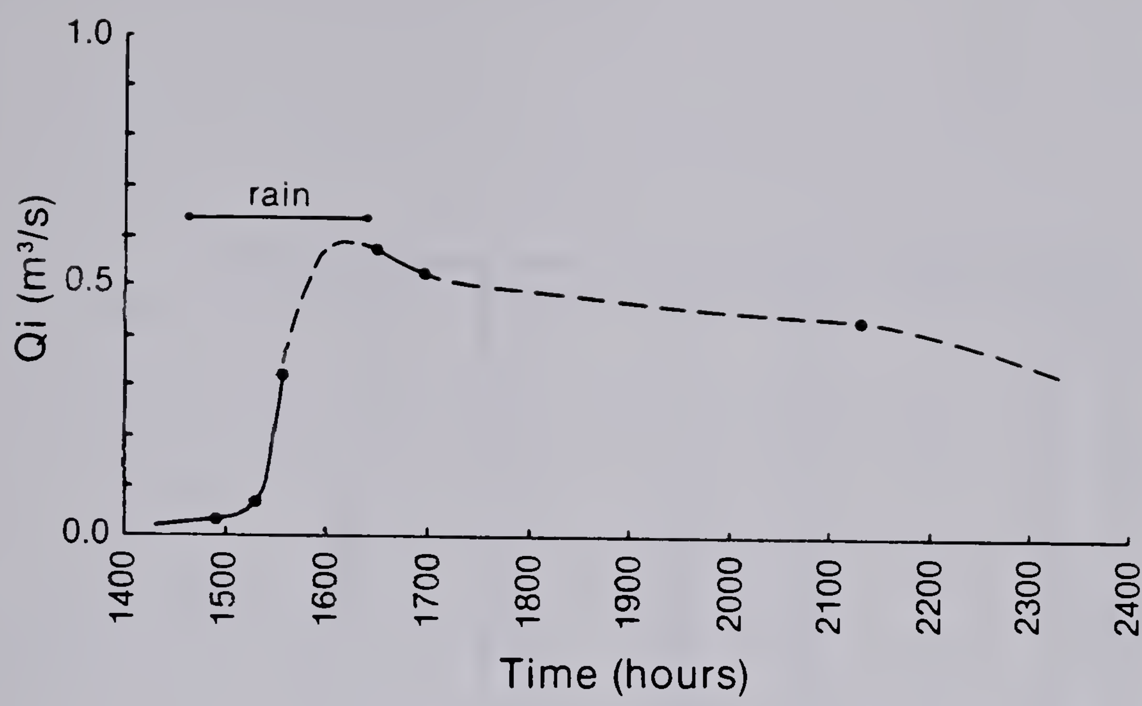


Figure 3.14 Rainstorm hydrograph, Mclean Creek, June 19, 1979

Kananaskis L.O., 2073 m, 32.90 mm; Moose Mountain L.O., 2434 m, 37.80 mm). Most of the rain fell on July 16 (67.8 mm at E.R.S., 28.4 at Kananaskis, 12.7 at Glenmore Dam).

The first response to the initial intense rainfall was a rapid rise to peak (45 minutes) and rapid fall to base flow one hour later at Falls gauge. The response at Bragg gauge was a similarly rapid rise an hour after Falls gauge and with a similarly rapid fall (Figure 3.15). These rapid rises and falls of the hydrograph occur frequently and probably represent the in - channel precipitation response to heavy local rainfall within the frontal system. These changes in discharge are not as common or as pronounced at Little Elbow River gauge (Figure 3.15).

The main hydrographic rise following the July 16 rain storm began on the 17th at 0010 hours at Bragg gauge and at 0045 hours at Falls gauge. The instantaneous discharge at

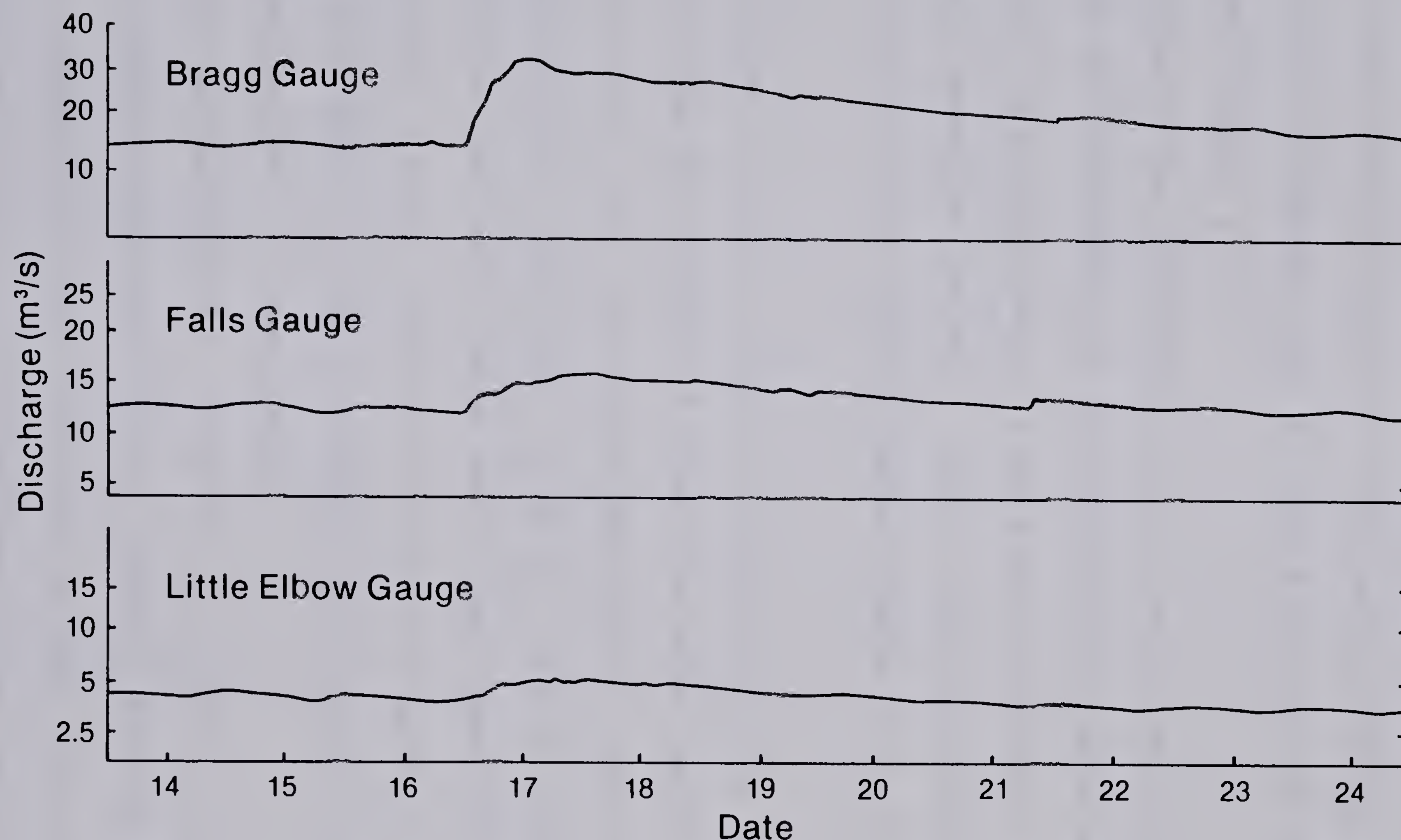


Figure 3.15 Hydrographs, Elbow River hydrometric stations, mid - July, 1978

Bragg gauge more than doubled, from 12.77 to 28.96 m^3/sec, to reach peak flow in 13 hours. It took 25 hours to reach the peak discharge at Falls Gauge and 24 hours at Little Elbow River gauge (Figure 3.15). The storm runoff peak daily discharge occurred the day after the rain at Bragg gauge and two days after the major rain at Falls and Little Elbow River gauge (Figure 3.16).

To evaluate the storm runoff of the different zones represented by the gauges, a flow recession curve was drawn through the lowest mean daily flow points for each station to represent baseflow (Kunkle, 1962), and the storm runoff, specific to each zone, was estimated (Table 3.6; Figure 3.16).

The volume of storm runoff in gross and unit terms is far greater from the foothills than the upper basin. The total storm runoff for the eleven day period July 16 to July 27 for each specific area was, in da m^3, 553 at Little Elbow River, 1822 for the Falls Gauge sub - basin and 4201 for the Bragg gauge sub - basin. The unit yield for each sub - area was 4.32, 5.93 and 11.7 mm of runoff, or 4.14, 5.73 and 11.37 litres/sec/km^2, respectively.

The pronounced spatial variation in runoff is probably due to the spatial distribution of rainfall and the greater hydrologic response of the foothills. As well, the location of the storm centre has an obvious bearing on the output from the various zones delineated by gauge location. As mentioned previously there is a considerable spatial

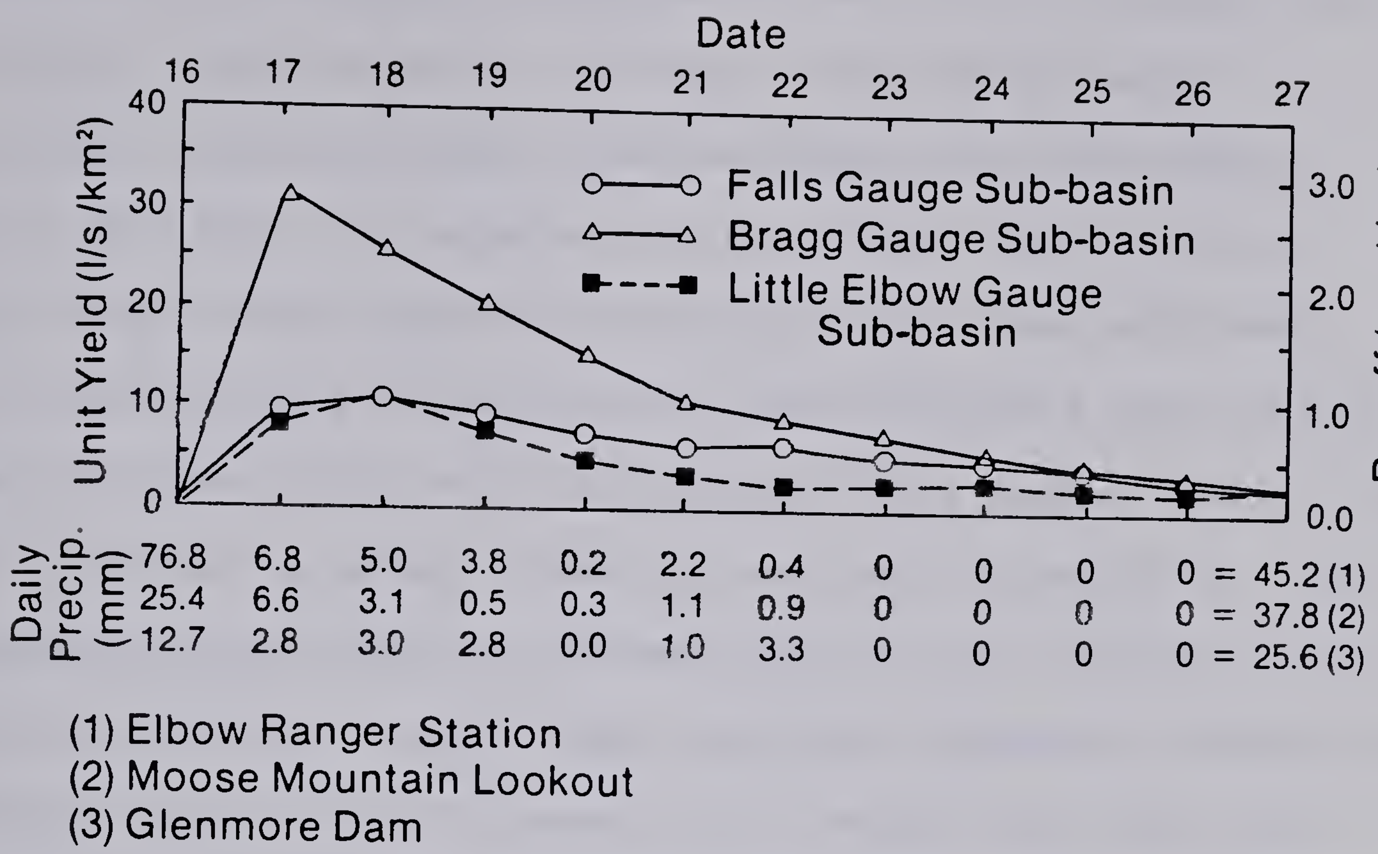


Figure 3.16 Unit area storm runoff and daily precipitation, mid and upper Elbow River basin hydrometric stations, July 1978

Table 3.6 Unit area storm runoff and daily precipitation, mid and upper Elbow River basin hydrometric stations, July 1978

Specific Mean Daily Storm Yield* (m^3/s)

Date	Little Elbow Gauge	Falls Gauge	Bragg Creek Gauge	Units
July 16	0.00	.14	.28	m^3/s^{-1}
17	1.27	2.52	11.41	
18	1.36	3.31	9.17	
19	.96	2.88	7.22	
20	.59	2.29	5.44	
21	.40	1.93	3.85	
22	.28	2.01	3.28	
23	.31	1.61	2.58	
24	.37	1.47	1.95	
25	.31	1.30	1.50	
26	.25	.91	1.22	
27	.28	.71	.99	
Mean	.53	1.76	4.07	m^3/s^{-1}
Mean Unit Yield	4.14	5.73	11.37	$l/s^{-1}/km^2$
Unit Peak Yield	10.63	10.78	31.87	$l/s^{-1}/km^2$
Total Yield	553	1822	4201	da m^3
Drainage Area	128	307	358	km^2

* Based on hydrograph separation to define base flow conditions following Kunkle (1962). Specific storm yield is the runoff at a station minus the contribution from baseflow and minus the inflow from upstream of the specific sub-basin.

variation in recorded precipitation which is not quantified in detail over the basin. Although the initial major rainstorm occurred almost simultaneously over the whole basin, the rate of response at Bragg gauge was far more rapid than at the upstream stations. Peak flow occurred in half the time and maximum storm runoff passed a day earlier than at Elbow Falls and Little Elbow River gauges. Runoff in the foothills occurred primarily as rapid throughflow to a perennial stream network whereas most of the mountain tributaries draining a large sub - area are ephemeral and had no surface runoff for this event. As a result the runoff for the mountains for this storm is about 7% of the estimated precipitation, whereas it is about 17% for the mid foothills.

The runoff value of 17% was derived from an estimated rainfall of 70mm in the mid foothills. This rainfall value, based on a Thiessen polygon approach (Raudkivi, 1979), is problematical given the topography of the area and the local complexity of rainfall in this environment (Longley, 1972; Appendix 2). In addition, runoff from the storm may be under - estimated because storm runoff would probably continue several more days into the next rainstorm, which would produce an estimated additional 910 da m^3 of delayed runoff. The runoff over 20 days would therefore be about 20% of the estimated precipitation input. Monenco (1979:B12) suggests that for the whole of the Elbow River basin above Bragg gauge the runoff for single storms is 25 to 60% of

precipitation, depending mainly on antecedent moisture conditions. Over the long term the runoff coefficient averages 51% of precipitation for the May to October period.[2]

The estimated inflow into Glenmore Reservoir in the 1935 to 1978 period is not considered to be accurate enough to meaningfully differentiate the hydrologic response of the plains from the remainder of the basin. There were no significant storms over the lower basin in 1979, when the river was continuously gauged. The increase in discharge downstream between Bragg gauge and Sarcee gauge recorded in 1979, resulted from groundwater inflows near the plains - foothills margin (Appendix 3). However, the runoff response of adjacent plains areas suggests that the lower Elbow River basin can produce significant runoff from heavy, stationary front rain storms (Appendix 2).

Late in the summer of 1978 and 1979, the Upper Elbow River infiltrated and flowed beneath the surface of the alluvium of the valley floor above the confluence with the Little Elbow River. The seepage loss was probably significant judging by the discharge of the major tributaries in the adjacent Little Elbow River basin (Table 3.5). For example, on August 29, 1979, the discharge recorded in the Upper Elbow River below Cougar Creek was 1.8 m^3/sec and further downstream, 3 km above the confluence with the Little Elbow

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[2]The average runoff is 281mm (based on 1967 to 1979 data) and the average precipitation at Elbow River Ranger Station is 551.8 mm (Table 2.5).

River, the discharge was 0.16 m^3/sec. Surface runoff ceased 0.5 km upstream of the confluence with the Little Elbow River (Stelfox and Konyenbelt, 1980).

Canyon Creek, which is the largest foothills tributary (121 km^2), is also ephemeral. The loss in discharge downstream (Stelfox and Konyenbelt, 1980) is due to infiltration into the stream bed (Borneuf, 1980). Although the stream is confined by bedrock in the lower reaches, the subterranean flow does not re - emerge at this location. It is postulated that the groundwater losses in Canyon Creek, and other foothills tributaries, re - emerge at the lower foothills - plains boundary and further downstream (Hitchon, 1969; MacMillan, 1980 ; Appendix 3). Swanson (1970) suggests that comparable subsurface inter - basin transfers occur in a similar geologic situation in the Porcupine Hills, which are immediately south of the Elbow River basin.

F. FALL RUNOFF

A transitional period in precipitation occurs around October. Summer convectional rains abate and as a result the hydrograph no longer reflects rainfall generated floods. The mean daily temperature is usually above zero until November. Precipitation is light and falls mainly as snow, which usually does not remain on the ground. There is little correspondence at this point between air temperature and discharge. However, in November and December discharge trends parallel temperature trends during this overall

receding flow period (e.g. Figure 3.8). The river usually freezes over within two weeks of November 10.

G. SIGNIFICANCE OF THE HYDROLOGIC REGIME

The significance of the hydrologic regime to sediment transport pertains to how and where runoff is generated, and magnitude and frequency characteristics of the flows which are generated.

The Elbow River has a typical high latitude runoff regime (Church, 1974). Winter discharges are low and the river and tributaries are frozen over for about five months. The major sources of winter runoff are groundwater influxes in the mountains and foothills. The plains area is generally in a water - deficit situation so that groundwater inputs are small. Winter precipitation occurs as snow, which remains on the ground for several months in the mountains and foothills. The storage as snow effectively redistributes several months precipitation into the short duration snow melt period. The plains snow pack is frequently removed previous to "spring" by warm, dry, chinooks which cause major sublimation losses and probably some surface runoff over the frozen ground. Depression storage and percolation may saturate the valley wall deposits and cause slumping into the river channel (Dawe, 1980).

The bulk of the annual runoff occurs as the result of snow melt. Maximum melt generally occurs in April in the plains, in May in the lower foothills and in June in the

upper foothills and mountains. The early spring floods are relatively small and occur during and after river ice break up and melt. The main flood events are generated by upper basin runoff in late May or June. By this time the river ice cover has degenerated so that only a few, small, isolated patches of ice occur along the river channel. However, this is not the case for high elevation, low order streams. These streams may flow over, under or through the snow and ice cover of the channels and hillslopes.

Smaller floods tend to be generated as the result of snow melt alone. As the magnitude of the flood increases, the rainfall component tends to increase and the contributing area tends to extend further eastwards. Large floods are generated when substantial rainfall occurs at the time of the spring melt. The whole drainage basin may contribute significantly to such a flood. This variation in the source area of flood generation suggests that the upland component of lower basin sediment yield can only be significant during large flood events. On the other hand, the foothills and mountains are significant source areas for any flood scenario.

The annual spring flood is a short - lived event. About 25% of the total annual runoff occurs in June and 60% occurs in May, June and July. Discharge increases rapidly from the low winter flows as a series of diurnal flood peaks, which increase in magnitude with time, culminating in a single, significantly larger peak. The diurnal fluctuation in

discharge is substantial. The annual maximum instantaneous discharge is usually 1.5 to 2.5 times larger than the corresponding mean daily discharge. In effect then, a flood peak is generated every 24 hours during the snow melt period. The pattern may be modified by rainfall. The recession from the spring flood peak is very rapid. The diurnal snow melt pattern persists through July as the snow zone retreats to higher elevations. The recession is characterized by several, largely rainfall generated, flood peaks, which are generally smaller than the spring peak. The competence of the system to remove sediment must be limited by the short duration of these high discharges.

Summer rainfall may generate moderate floods. This rainfall, which occurs as intense, local showers or wide-spread frontal rain, generates a rapid throughflow response from the mountain and foothill streams. The plains area is generally in a water deficit during the summer. As a result there is generally little surface runoff, and much of this is captured by internally drained areas. Heavy frontal rain may produce substantial runoff from the plains area. Slumping of the valley wall deposits was observed to occur after periods of long duration, low intensity rainfall. The only areas which were observed to produce infiltration - excess overland flow were denuded seismic trails and road margins, particularly in the foothills. Gullies in the mountains and foothills appear to be initiated from slumping and appear to be maintained by throughflow interception and

snow melt. As a result upland contributions to sediment yield would be expected to be small. The subsurface translation of precipitation as throughflow would be expected to result in significant dissolution of rock and regolith. In addition to these short distance, rapid, throughflow systems, a deeper transient groundwater system has been identified. Deep percolation and channel seepage in the mountains and foothills is effluent at the plains - foothills boundary. As well, regional flow systems are effluent in the lower basin. These influxes are identifiable, in part, by high solute concentrations (Appendix 3).

4. THE SUSPENDED SEDIMENT REGIME

A. METHODOLOGY

Suspended sediment concentrations were measured at 18 sites in the Elbow River basin (Figure 4.1) using USDH 48 and 59 samplers at representative verticals, in rated sections, by the equal transit rate method (Guy and Norman, 1970) (Photo 4.1). Daily or more frequent samples were taken during the spring flood period in 1978 and 1979 and miscellaneous samples were collected for the remainder of the summer. Water Survey of Canada systematically collected samples from the Elbow River at Bragg gauge (station 5BJ4), to compute daily suspended sediment concentration and loads during the March to October period, and one winter period, for six years between 1969 and 1977.

A modified ISCO 1680 waste water sampler was used to automatically collect suspended sediment samples at Sarcee gauge (Figure 4.1). Suspended sediment samples were collected by the ISCO at 3 to 8 hour intervals from mid May until mid July and at miscellaneous intervals each day until September in 1978. In 1979 samples were collected at 3 to 8 hour intervals from May through August. During the summer at least one sample per day was retained for analysis. If there was perceptible sediment in the sample bottles, those bottles and proximal samples were retained for analysis.

The ISCO, which is a self purging, peristaltic pump sampler, was fitted with a 6.35 mm USDH 48 sampler nozzle,

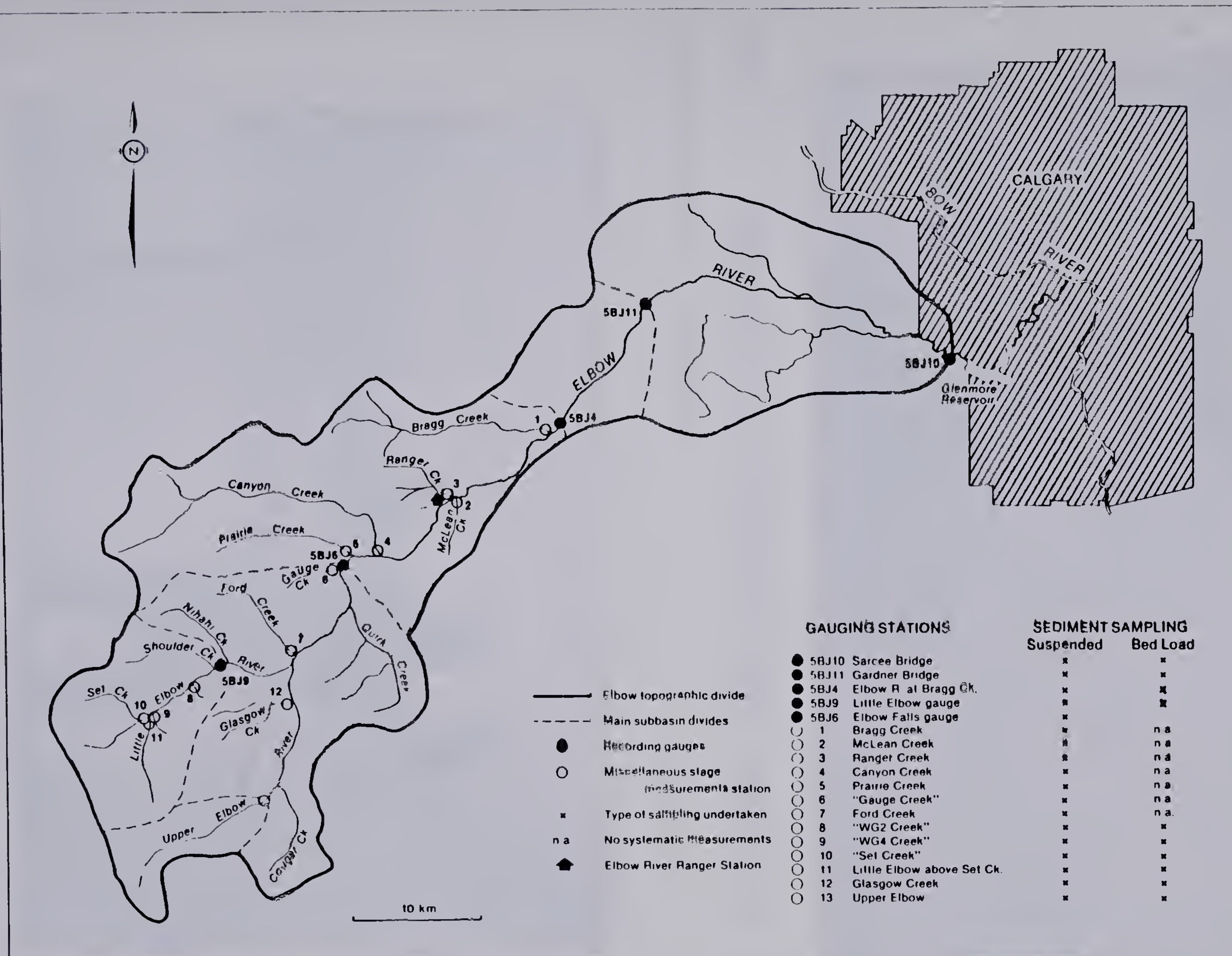


Figure 4.1 Elbow River basin hydrometric stations and sediment sampling stations

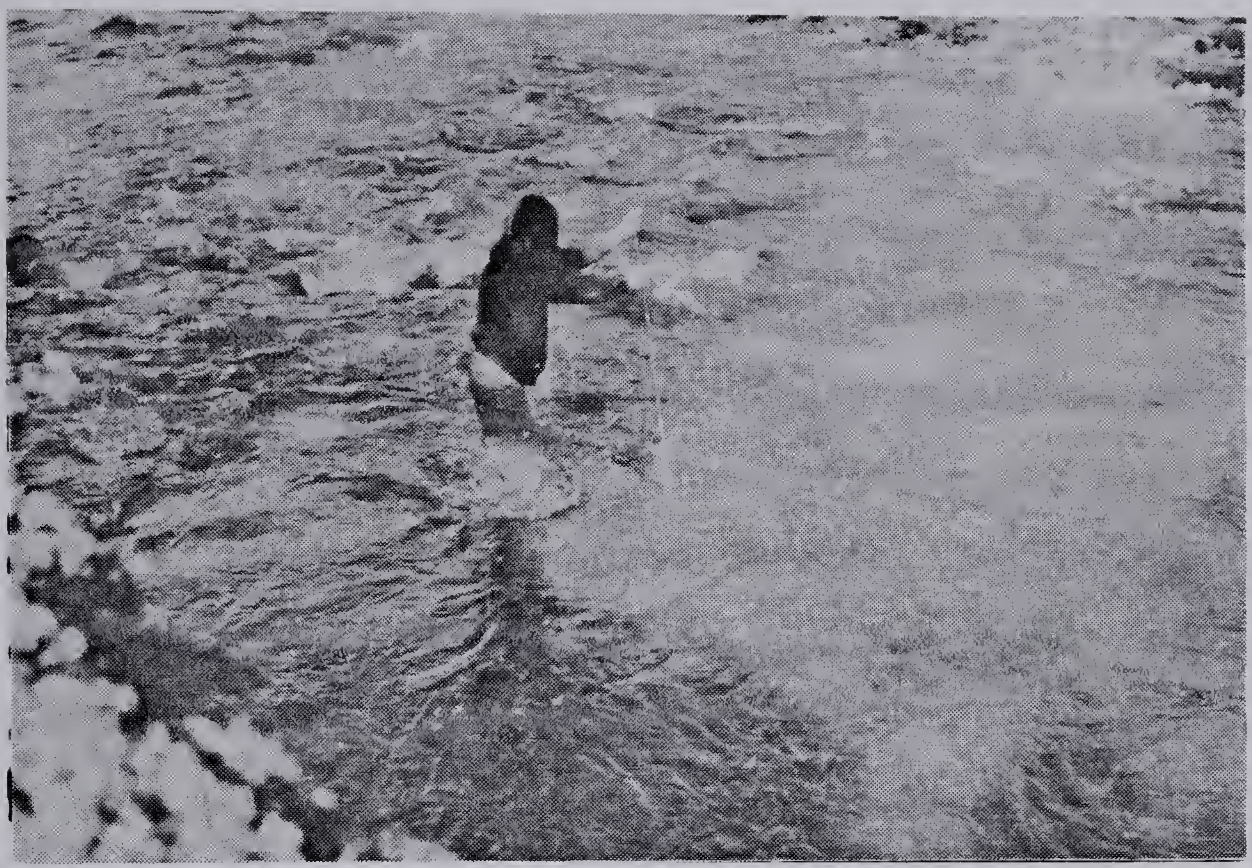


Photo 4.1 Sampling with a USDH 48 water sampler at a representative vertical

Photo 4.2 The ISCO automatic sampler was fitted with a USDH 48 sampler nozzle which was projected horizontally into the flow

which was projected horizontally into the flow 20 cm above the bed 3 m from the bank (Photo 4.2; Figure 4.2). The flow depth ranged from 40 to 80 cm over the measurement period. The sampler was stationed about 90 m downstream of Sarcee gauge, at a stable cross - over. The site is not affected by tributary inflow and is far enough downstream of active bank erosion sites for mixing to occur. Immediately before taking each 500 ml sample, the sampler line was flushed. Samples are pumped by contraction of the pipe not by an impeller. Thus, when the intake tube was taut, and sloping directly to the intake nozzle, cross - sample contamination should not have occurred.

One extensive calibration was undertaken at a high discharge at the ISCO site, and several miscellaneous replicate samples were taken at a representative point from Sarcee Bridge using USDH 48 and 49 samplers (Figures 4.2 and 4.3). The calibration information suggests that the intake location is representative of the mean cross - sectional suspended sediment concentration (load / discharge) and that the ISCO obtained representative samples.

There is some scatter in the relationship between the ISCO and replicate USDH 48 and 59 samples (Figure 4.3). However, the calibration data plot as an almost 45 degree line. Thus, there appears to be no high or low bias, at least under the test conditions. Paustin and Berchta (1979) found that there was no significant difference in suspended sediment concentrations in samples from a US DH 48 sampler

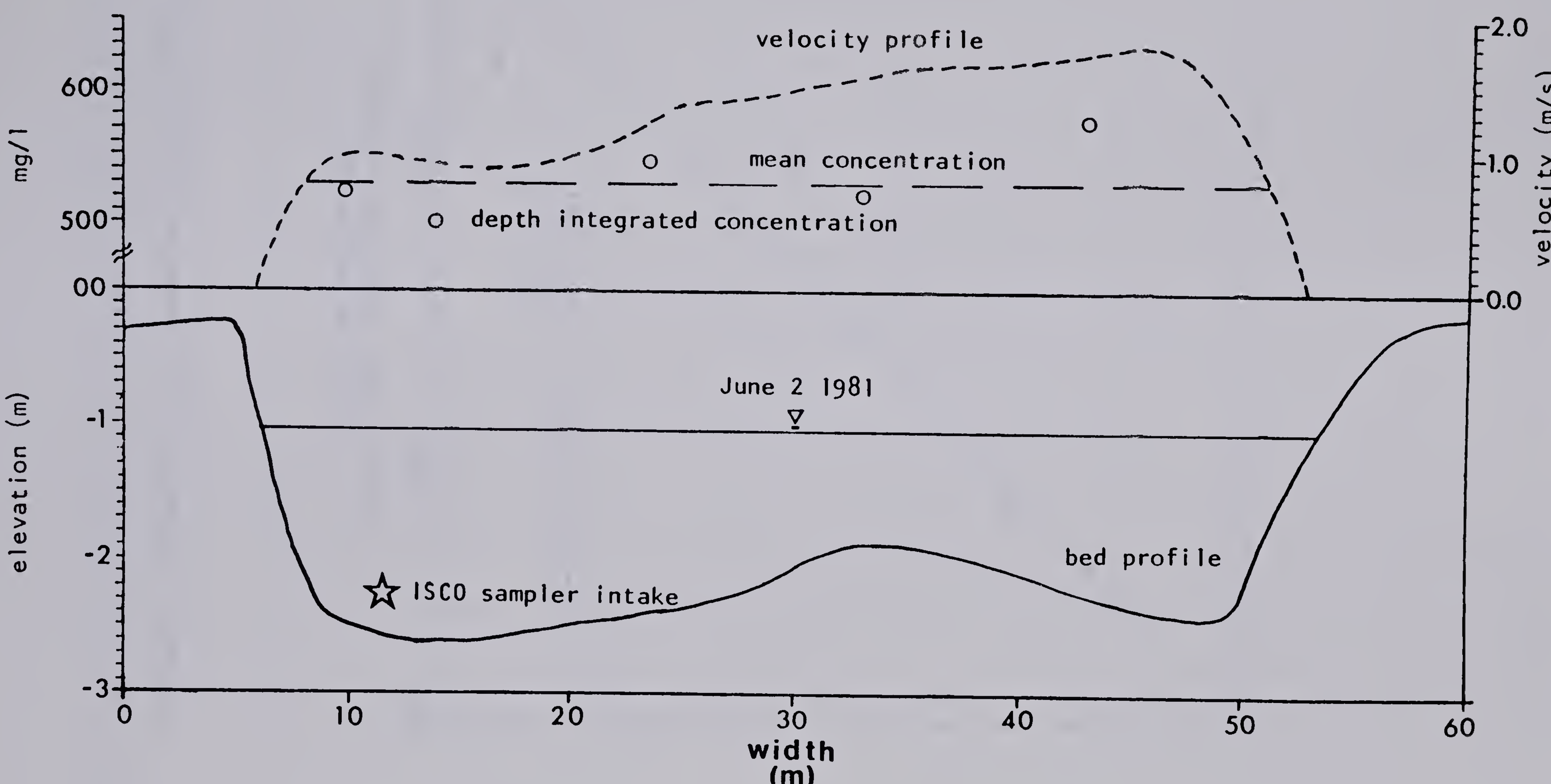


Figure 4.2 Hydraulic geometry and suspended sediment concentrations at the ISCO sampling site near Sarcee gauge

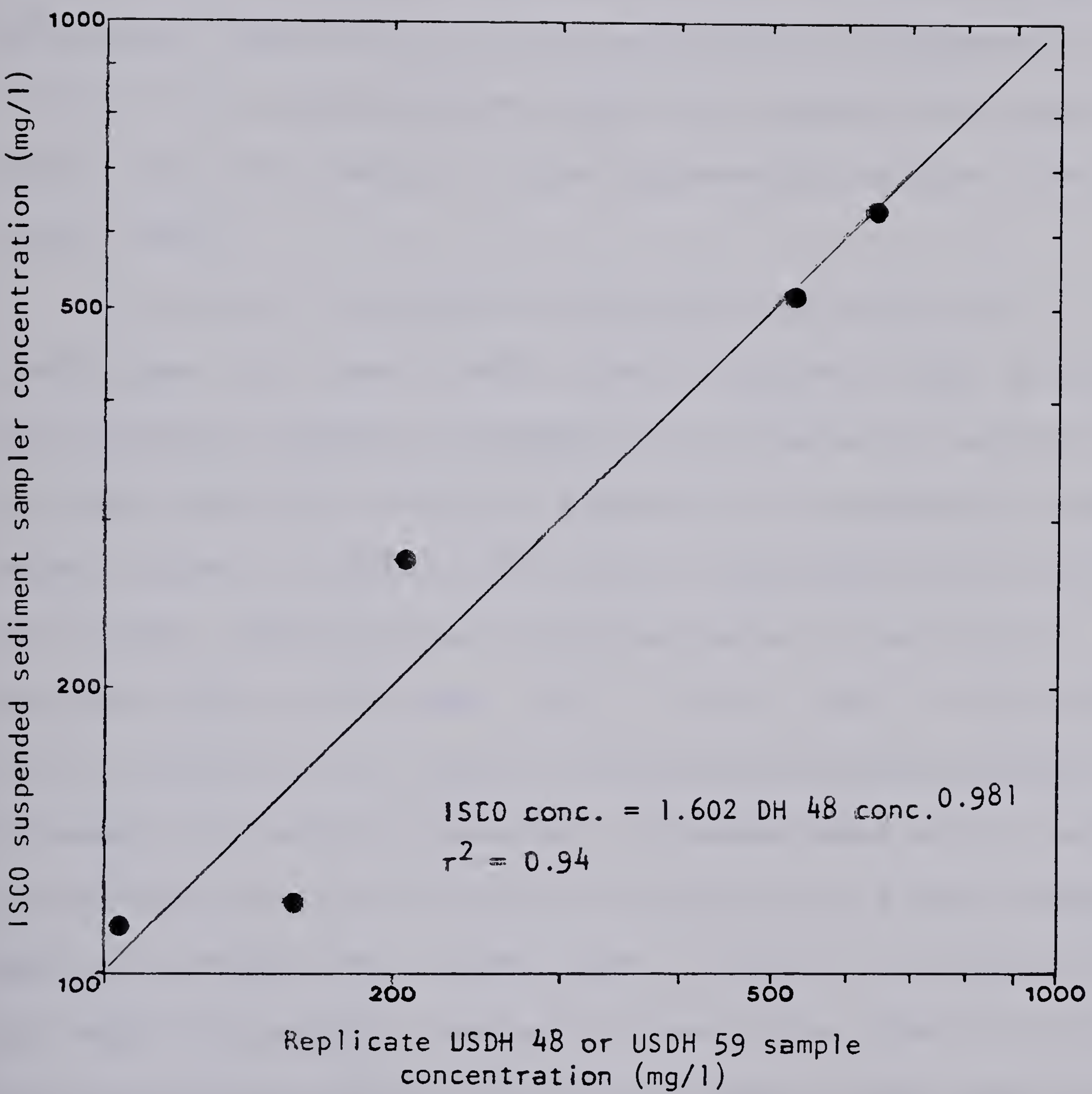


Figure 4.3 Relationship between ISCO automatic sampler suspended sediment concentrations and replicate US DH 48 and 59 samples

and an ISCO sampler, in a similar gravel bed river, for concentrations less than 54 mg/l. They found the sampler slightly over - sampled at higher concentrations. As a result of the calibration tests no correction for the sampler performance, or sampler location, was considered necessary. However, it is recognised that considerable variation in concentration over short periods may occur, which may mean samples are not representative over time (Guy, 1970),

In terms of standard measurement and analytical techniques (e.g. Guy, 1969) fluvial sediment load is divided into material moving in suspension and material moving near the bed. Material moving in suspension is sampled by suspended sediment samplers. The wash, suspended bed material and solute components are differentiated by laboratory analysis (Guy, 1970; Hem, 1970). Solute load is dissolved rock in solution. As such it is clearly distinguishable from movement of clastics. However, in established analytical procedures dissolved material in solution and rock fragments may be confused. The solute load is usually considered to be the material passing through a given filter. The filtrate may exclude some dissolved material (Hem, 1970), but may include fine clastic material, which is wash load (Douglas, 1971).

The suspended sediment load of the Elbow River at Bragg gauge is composed almost exclusively of particles less than 1.0 mm, at least for discharges up to about 100 m^3/s, a four

year flood (Figure 4.4). The Elbow River has a bouldery thalweg and gravel bars which contain only a small, but variable, proportion of fines. Given the definition in Chapter 1, the Elbow River suspended sediment load is wash load, because the load is "...composed of particles sizes finer than those represented in the bed and is determined by available bank and upslope supply ...." (Simons and Senturk, 1976:506).

Suspended sediment discharge was estimated from rating curves based on known or estimated discharge (Porterfield, 1971). The rating curves were developed for short periods because the suspended sediment concentration - discharge relationship underwent significant shifts. The nature of these shifts is discussed in following sections.

B. ANNUAL LOAD ESTIMATES

A major purpose of the research was to estimate the contribution of the mountains, foothills, and plains to the long - term (1932 to present) sedimentation in Glenmore Reservoir (Figure 4.1). Temporal variations in load are treated in detail in the following section. The long - term loads at Bragg gauge could be estimated from eight years of data. Missing winter and summer load totals were estimated from two models based on 60 months of record (Table 4.1). The rising limb monthly load - monthly runoff log model, which is based on 30 months of data, explains 94% of the suspended sediment load for the months up to and including

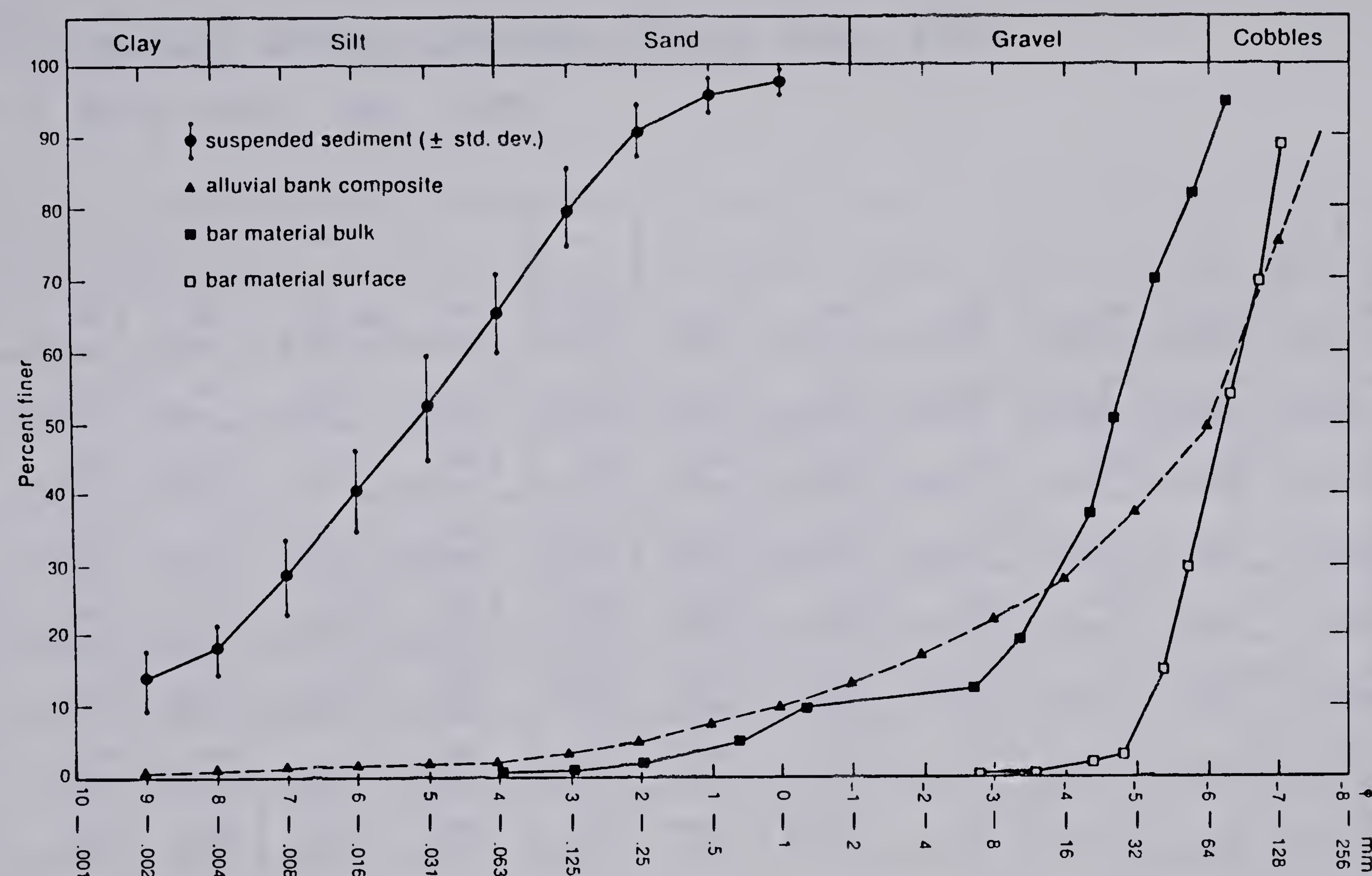


Figure 4.4 Bulk sample particle size distribution, Elbow River reach near Bragg Creek

Table 4.1 Measured monthly suspended sediment loads, Elbow River at Bragg gauge, 1969 - 1979

YEAR	J	F	M	A	M	J	J	A	S	O	N	D
1979	8E 5811	9E 6210	32.2E 8962	77.7 12562	1447* 35544	122 33877	59.6 21595	17E 15639	13E 12693	10E 10495	8E 8131	8E 8303
1978	11E 6558	6E 5423	25E 8838	20.6E 12602	753 36136	5714* 60541	594 42331	60.5 25403	32E 21138	22E 17738	12E 12561	8E 8086
1975	6E 5410E	18E 7926E	19.0 8301	52.2 10344	226 24960	4836* 52949	811 40552	21.5 22631	48.9 15377	39.9 13248	10E 10877E	9E 9242E
1974	14E 7144E	7E 5771E	47.5 8920	564 16020	2435 54245	30938* 92507	247 38257	87.4 28791	33.4 22650	19.1 17075	14E 13836E	9E 10051E
1973	16E 7548E	15E 7435E	16.0 8974	44.5 12732	4009 53705	3266* 66843	99.5 32938	12.3 20069	40.3 21234	17.7 14149	11E 11565E	9E 9490E
1972	15.8 7411	11.7 6344	84.4 10909	96.1 13109	4180 49230	9645* 81204	162 44682	63.2 27974	15.5 20091	49.4 16634	14E 13609E	10E 10352E
1971	14E 7142E	10E 6454E	5.1 7219	228 14746	1215 45122	17049* 7576[illegible]	82.8 30166	61.6 18655	19.8 14131	0.0 13525	41.2 11262	41.3 8492
1969	28E 9194E	15E 7444E	20E 8142	284E 21050	1975 56677	4631[illegible]* 85978	12523 75494	18.5 20546	6.3 17322	0.0 15154	13E 13135E	10E 10380E
load $\bar{x}$ s	14.1 6.7	11.5 4.2	31.2 24.8	171 184	2030 1444	14735 16104	1822 4332	42.8 28.6	26.2 14.8	19.8 17.6	15.4 10.6	13.0 11.4
Q $\bar{x}$	7027	6626	8783	14146	44452	68708	40752	22464	18080	14752	11872	9300

E Estimated load

the peak runoff month (Figure 4.5). A semi - log model explains 68% of the monthly suspended sediment load after the peak runoff month (Figure 4.6). The two measured zero load months (October 1969 and October 1971) were arbitarily assigned a value of 1.0 tonne per month. This results in an almost identical curve to that for the monthly load - monthly discharge relationship ($r^2 = 0.90$) which exists for monthly discharges greater than about 25,000 da m^3, when most of the load is carried. The monthly sediment load - runoff relationship for discharges less than 25,000 da m^3 is almost random ($r^2 = 0.02$).

The estimated average suspended sediment load at Bragg gauge, for the period of hydrometric records 1935 to 1979, is 18,186 t/yr (Table 4.2). For the eight years of measured data, the average load is 18,932 t/yr (Table 4.1).

The long - term average annual suspended sediment load of the upstream station was estimated from the relationship between measured loads at the stations and at Bragg gauge for coincident periods in 1978 and 1979. Assumptions were made to derive the long - term loads from the upper basin stations. Because loads were measured for only the spring and summer periods, the assumption was made that the 1978 and 1979 annual loads at Bragg gauge could be reasonably estimated using the ratios of May to August loads on March to October loads and annual loads of previous measured years at Bragg gauge. Further, the assumption was made that these ratios were maintained at the upstream stations. The annual

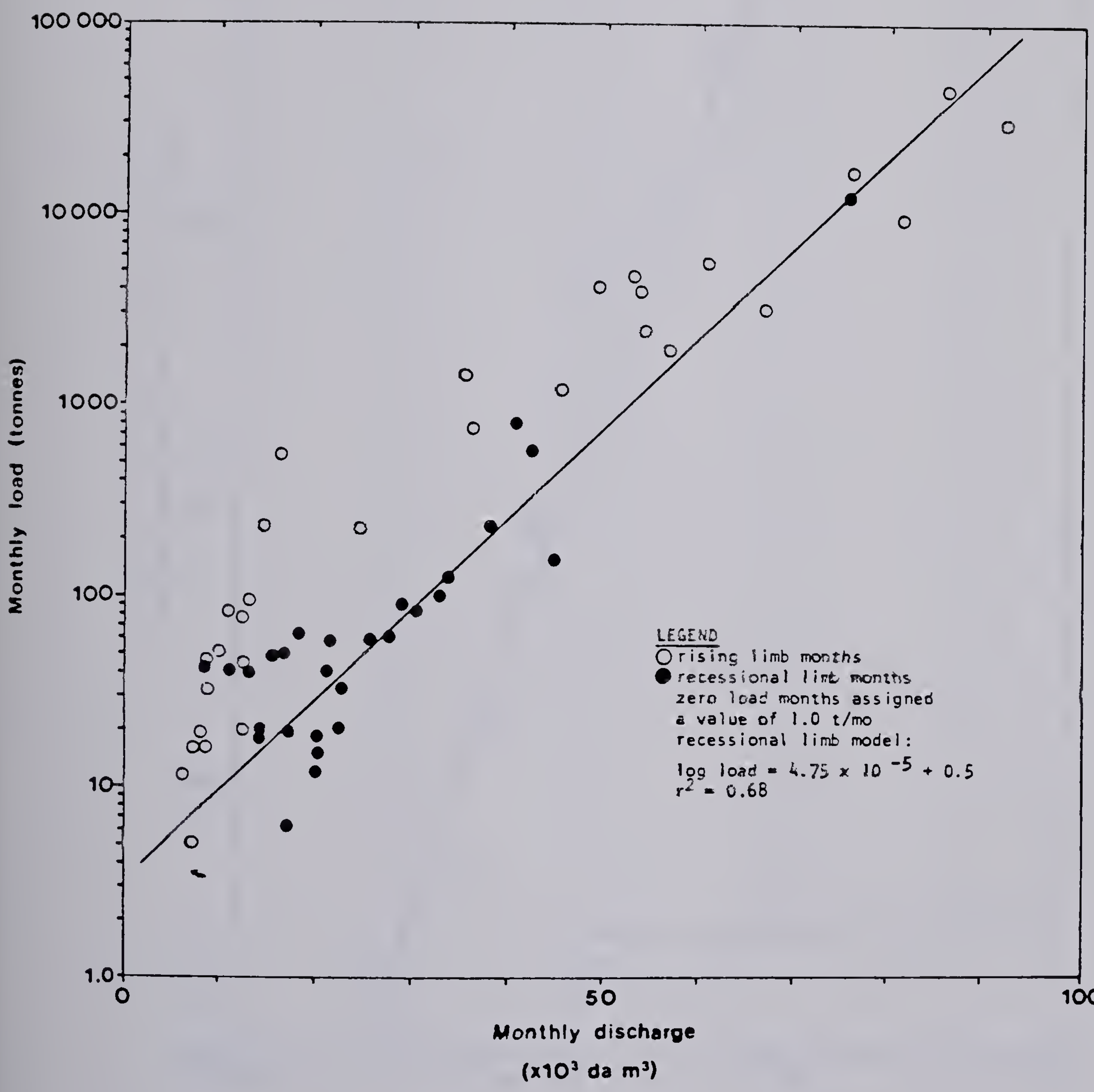


Figure 4.5 Measured monthly suspended sediment loads against discharge for rising limb months, Elbow River at Bragg gauge

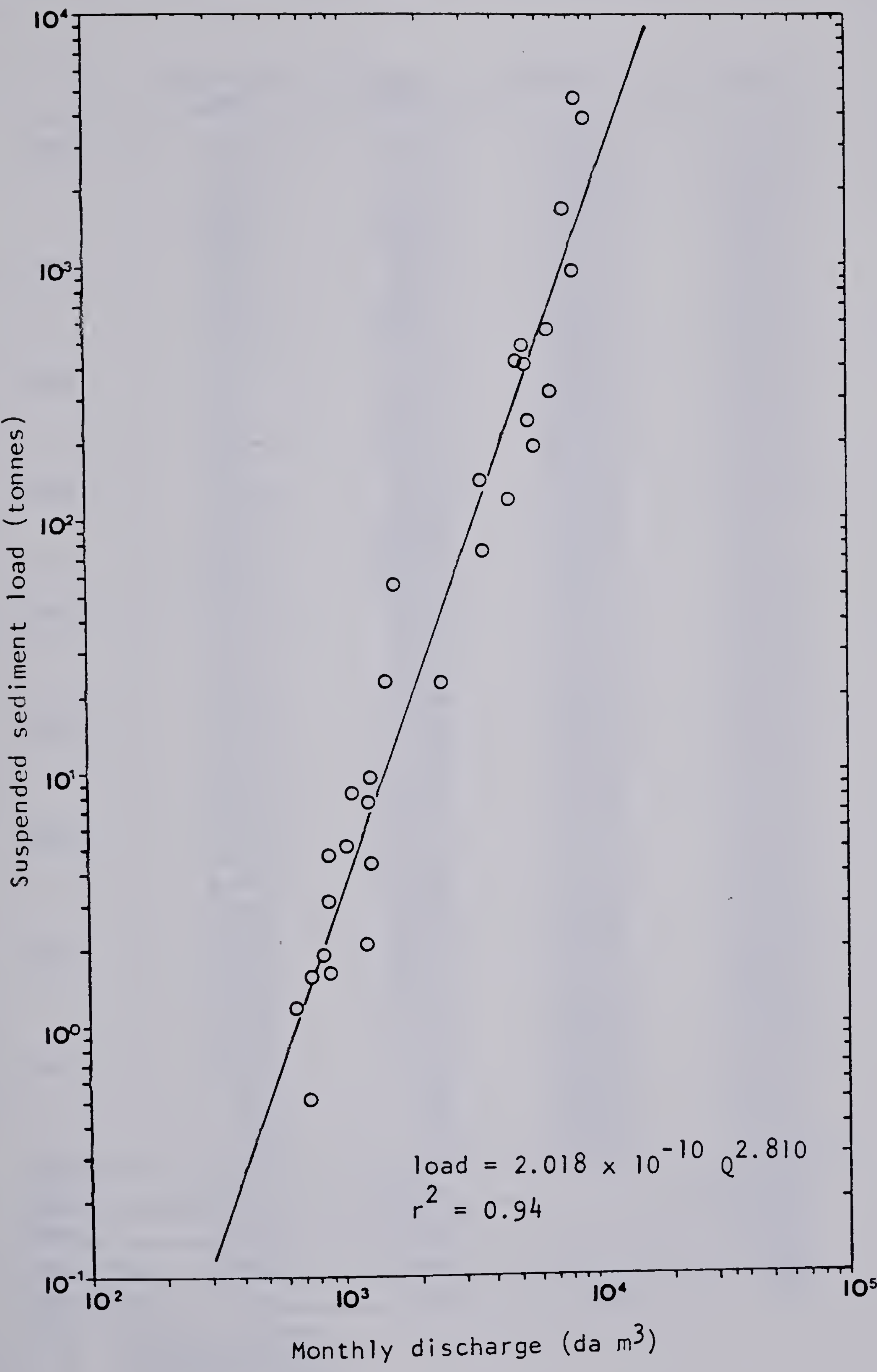


Figure 4.6 Measured monthly suspended sediment load against monthly discharge, Elbow River at Bragg gauge, 1969 - 1979

Table 4.2 Estimated annual, and March to October, suspended sediment loads, Elbow River at Bragg gauge, 1935 - 1979

	Mar-Oct load	Annual load	Mar-Oct Q	Annual Q
1935	1848	1877	150353	177736
	320	338	83184	104433
	4104	4138	161384	195675
	22878	22920	274770	308250
	18203	18250	205657	244521
1940	1730	1769	174052	209496
	2532	2571	142507	176090
	71298	71351	456596	499931
	28234	28301	292097	329652
	694	725	147401	176874
1945	53736	53784	350816	392222
	6360	6435	235990	281994
	18416	18508	314609	367950
	50518	50225	363489	404556
	1282	1328	138284	171093
1950	5438	5475	191476	222889
	73732	73882	468672	528659
	14097	14228	295564	347229
	96640	96703	368873	409706
	10567	10660	282467	337508
1955	8029	8094	224848	262727
	5500	5546	192964	226249
	4713	4756	182287	219862
	16487	16531	264686	300503
	5145	5189	194064	231725
1960	3458	3500	176233	207650
	7257	7300	220676	258484
	3936	3989	181380	213981
	11661	11701	223589	257224
	18916	18955	240812	274170
1965	17589	17647	303047	350083
	6429	6494	218162	256646
	49535	49568	312626	342662
	5500	5543	211740	248708
	61138	61204	300363	340516m
1970	13236	13283	290046	242203
	18662	18768	219325	252675m
	14295	14347	263833	301549m
	7506	7557	230644	266682m
	34371	34415	278465	315267m
1975	6055	6098	188362	221817m
	4441	4483	186980	222428
	827	872	130203	164273
	7222	7259	224727	257355m
1979	1779	1812	151367	179822m
Total	816314	818379		12301725
mean	18140	18186		273371
Std. dev.	22630	22628		86871
Coef. var.	125%	124%	34%	32%

N.B.: Load in tonnes
Q Discharge in da m^3
m measured loads

data from Bragg gauge suggests the spring and summer period (May to August inclusive) produces about 0.91 (1979) and 0.98 (1978) of the annual load. March to October inclusive loads total, on average, 99.4% of the estimated annual load. The respective reciprocals were used to estimate the 1978 and 1979 annual loads of the upstream stations based on the measured spring and summer loads.

The long - term loads of the sediment sampling stations upstream of Bragg gauge were estimated by multiplying the long - term average annual load at Bragg gauge with the ratio of the 1978 and 1979 loads of the stations on the coincident Bragg gauge load. The variance in ratios in 1978 and 1979 produce a range of values, for each of the sampling stations, which are reasonably limited (Table 4.3).

The mountains of the Elbow River basin produce in the order of 1800 to 2400 t/yr of suspended sediment, over the long - term. This translates into a long - term yield of about 11 $t/km^2/yr$. There is considerable variation in the yield for the mountain sub - basins (3.8 to 23.9 $t/km^2/yr$). About 75 to 90% of the load in the Little Elbow River basin is derived from tributary sources (Table 4.3).

The annual suspended sediment budget of the Fall's gauge sub - basin is summarized in Table 4.3. The loads from unmeasured tributaries were estimated, based on drainage area, from known loads in adjacent basins (Figure 4.7). About one third of the sub - basin area is floodplain, low terrace or valley wall with insignificant or no observed

Table 4.3 Estimated suspended sediment loads, Elbow River hydrometric stations

Little Elbow basin

Station	area (km^2)	Sediment load (tonnes) May & June 1978	Sediment load (tonnes) May & June 1979	Sediment load (tonnes) annual est. 1978	Sediment load (tonnes) annual est. 1979	Sediment load (tonnes) long term average	Sediment yield ($tonnes/km^2$) annual 1978	Sediment yield ($tonnes/km^2$) annual 1979	Sediment yield ($tonnes/km^2$) long term average
Upper Little Elbow	16.3	138	22	155	25	245 - 390	9.51	1.53	15 - 23.9
Set Creek	53.2	183	48	205	55	515 - 550	3.85	1.03	9.7 - 10.3
W. Glasgow 4	10.9	42	4.4	47	5.1	50 - 120	4.31	0.47	4.6 - 11.0
W. Glasgow 2	3.9	17	1.3	19	1.5	15 - 48	4.87	0.38	3.8 - 12.3
Sum of tributaries	84.3	380	75.7	426	86.6				
Little Elbow gauge	128	510	85	570	98	985 - 1430	4.45	0.77	7.7 - 11.2
non trib. sources	43.7	130	9.3	144	11.4	115 - 360			
non trib. %	34	25.5	10.9	25.3	11.6				

Falls gauge sub-basin

Station	area (km^2)	May & June 1978	May & June 1979	annual est. 1978	annual est. 1979	long term average	annual 1978	annual 1979	long term average
m Little Elbow gauge	128	510	85	570	98	985 - 1430	4.45	0.77	7.7 - 11.2
m Upper Elbow	65.3	357	69	400	79	790 - 1000	6.13	1.21	12.1 - 15.3
+ Cougar Creek	29.6	146	23	164	26	260 - 410	5.54	0.88	8.8 - 13.9
Ford Creek	21.1	103	22m	115	25	250 - 290	5.45	1.18	11.8 - 13.7
Gauge Creek	8.6	41	5m	46	6	60 - 115	5.35	0.70	7.0 - 13.4
+ Nahahi Creek	28.7	142	22	(159)	(25)	250 - 400	5.54	0.87	8.7 - 13.9
+ Quirk Creek	46.7	234	40	262	46	460 - 650	5.61	0.99	9.9 - 13.9
+ South Glasgow Ck.	11.6	55	7	62	8	80 - 155	5.34	0.69	6.9 - 13.4
+ Unnamed km 91.0	4.5	21	2	24	2	20 - 60	5.33	0.44	4.4 - 13.3
+ Unnamed km 98.0	8.0	38	4	42	5	50 - 105	5.25	0.63	6.3 - 13.8
+ Unnamed U.E. km 102.6	12.8	61	8	68	9	90 - 170	5.31	0.70	7.0 - 13.3
Sum of tributaries	365	1708	280		321	3215 - 4790			
Falls gauge sub-basin	242			1666	371	3725 - 4170	6.88	1.53	15.4 - 17.2
m Falls gauge	435	2354	477		548	5500 - 6600	5.41	1.10	12.6 - 15.2
balance	70	646	197		227	2285 - 1810			
non trib. sources %	16	27	41		41	42 - 27			

Bragg gauge sub-basin

Station	area (km^2)	May & June 1978	May & June 1979	annual est. 1978	annual est. 1979	long term average	annual 1978	annual 1979	long term average
Measured									
Bragg Creek	55.0	79.0	44.0	88.5	50.6	220 - 508	1.61	0.92	4.0 - 9.2
Canyon Creek	120.6	29.3	12.6	32.8	14.5	82 - 145	0.27	0.12	0.7 -
McLean Creek	31.6	53.5	22.4	98.0	71.7	245 - 720	3.10	2.27	7.7 - 22.8
Prairie Creek	41.8	33.6	18.6	37.6	21.4	94 - 215	0.90	0.51	2.2 - 5.1
Ranger Creek	20.1	67.0	27.3	75.1	31.4	188 - 315	3.74	1.56	9.4 - 15.7
Unmeasured									
Iron	12.7	3.2	1.8	3.5	2.0	9 - 20	0.28	0.16	0.7 - 1.6
Powderface	7.8	6.4	3.5	7.1	4.1	18 - 41	0.91	0.53	2.3 - 5.3
Silvester	14.9	25	11	46	34	115 - 340	3.09	2.28	7.7 - 22.8
Sum of tributaries	304.5	297	141	343	230	860 - 2308	1.13	0.76	2.8 - 7.6
sub-basin net	358.0	4113	1092	4623	1264	11582 - 12686	12.91	3.53	32.4 - 35.4
balance	53.5	3816	951	4280	1034	10723 - 12686	80.00	19.33	194 - 200

Sarcee gauge sub-basin

Station	area (km^2)	May & June 1978	May & June 1979	annual est. 1978	annual est. 1979	long term average	annual 1978	annual 1979	long term average
Sarcee gauge	1181			6711	8670	35000 -105000			29.6 - 88.9
*Reservoir	1210					75600			62.5

Silvester area weighted against McLean Creek
Powderface area weighted against Prairie Creek
Iron Creek, below the irrigation dam, area weighted against Bragg Creek
+Predicted load based on load-area model of 8 upper basin stations
m Measured loads
* Sedimentation survey

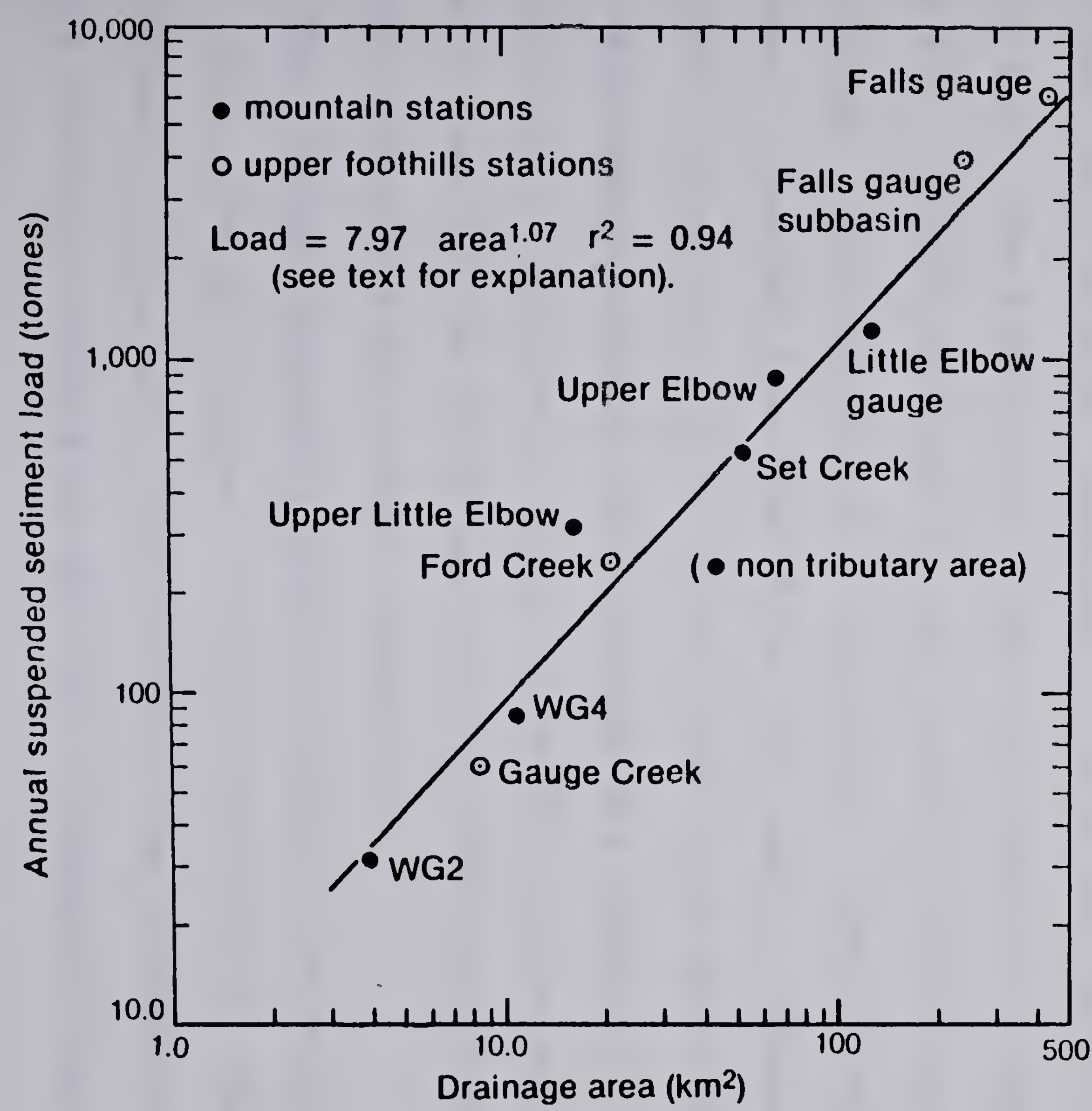


Figure 4.7 Long term average annual suspended sediment load against drainage area, Elbow basin mountains and upper

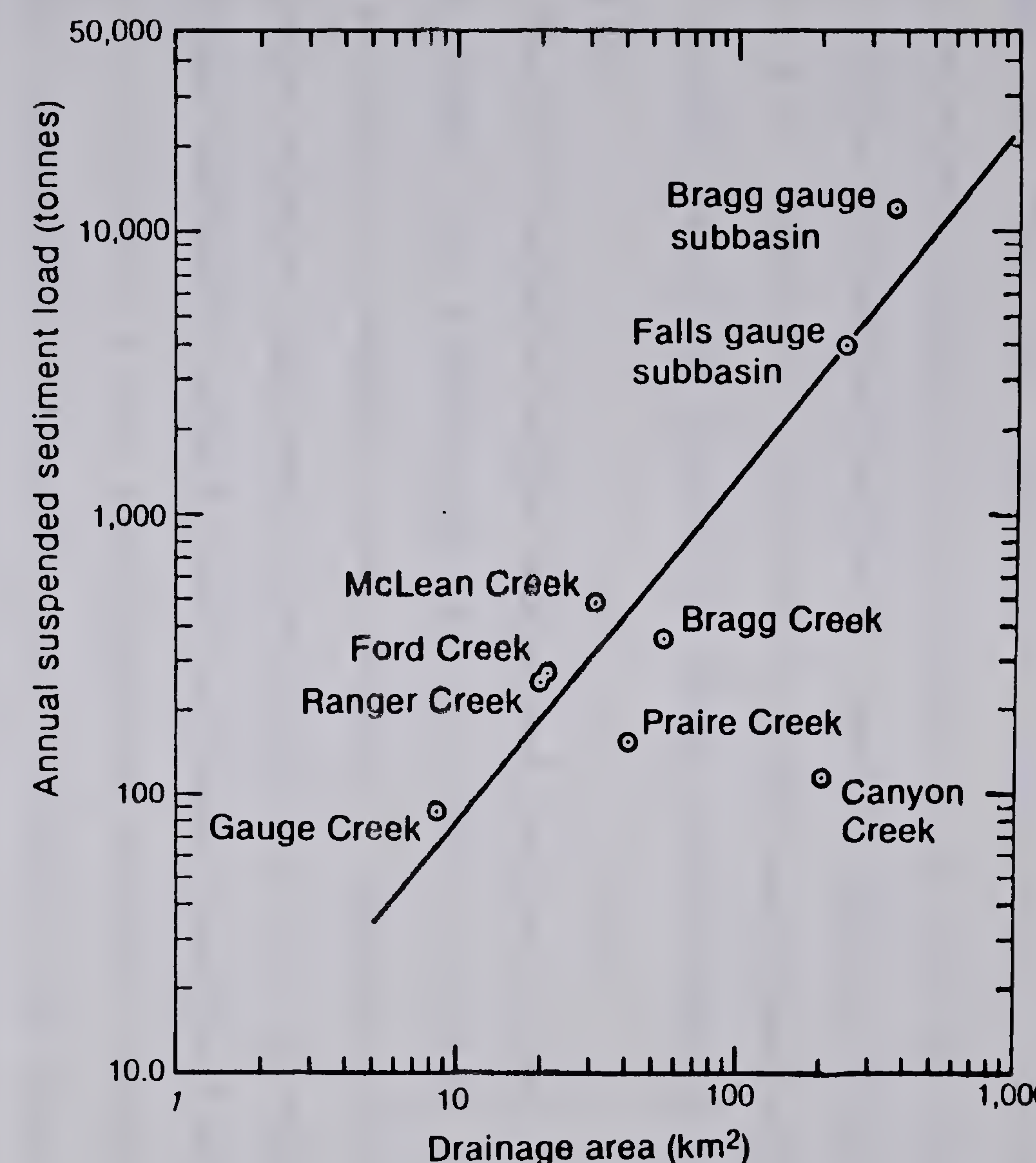


Figure 4.8 Long term average annual suspended sediment load against drainage area, Elbow basin foothills

streamflow. The major tributaries cover 71% of the sub - basin area and produce an estimated 39 to 57% of the sub - basin annual suspended sediment load of 5500 to 6600 t/yr (Table 4.3). That is, 43 to 61% of the suspended sediment load is derived from the river reach in the Fall's gauge sub - basin.

The estimated average suspended sediment load at Bragg Creek gauge, for the period 1935 to 1979, is 18186 t/yr. This estimate is based on the monthly - load monthly - discharge models discussed previously. For the 8 years of suspended sediment records the estimated average load is 18932 t/yr (Table 4.2). The suspended sediment load of the larger unmeasured tributaries was estimated, by area - weighting, with similar measured tributaries (Table 4.3 and Figure 4.8). The remaining 53 km^2 (or 15%) of the Bragg gauge sub - basin consists of very small tributaries, valley side slopes, low terraces and floodplain. Some suspended sediment is probably derived from very small, unmeasured, tributary streams, such as Connop Creek (3.8 km^2). However, the loads contributed by such streams are thought to be insignificant. Therefore, about 82 to 93% of the suspended sediment load of the Bragg gauge sub - basin would be derived from the river reach (Table 4.3).

The average suspended sediment load of the Bragg gauge sub - basin, in unit terms, is about 34 $t/km^2/yr$. There is considerable variation in the yield of the tributaries (0.7 to 22.8 $t/km^2/yr$). The 53.5 km^2 non - tributary area, which

represents 15% of the sub - basin area, has a unit yield of about 200 t/km²/yr.

The estimated suspended sediment load at Sarcee gauge was about 10,500 t in 1979 and 14,000 t in 1978 (Table 4.3). The estimated average long - term load at Sarcee gauge, using the ratio with Bragg gauge loads against the Bragg gauge long - term load, is 70,100 t/yr. The range in loads is considerable: 35,000 to 105,200 t/yr (Table 4.3). This estimate may be compared with annual load estimates from reservoir sedimentation surveys.

Hollingshead (1969) undertook a detailed survey of sedimentation in Glenmore Reservoir and concluded that the average rate of infilling of the reservoir was 61 acre feet/yr, which is about 75,250 m³/yr. The average specific gravity of the reservoir sediment was estimated, from tables, to be 1.20 (75 lb/ft³; Hollingshead, 1969). The estimate of specific gravity has been revised to 1.005, based on five sediment cores. Thus, the average reservoir deposition would be 75,626 t/yr. The average composition of the reservoir delta area is 35% sand, 50% silt and 15% clay (Hollingshead, 1969). The extent of gravel deposits in the reservoir has not been quantified, but is thought to be insignificant (Hollingshead et al., 1973). In Chapter 5, average bedload input is estimated to be approximately 250 t/yr.

The reservoir was resurveyed in 1977, by the City of Calgary, and this was supplemented by a survey of the upper

delta in 1982 (Photo 4.3). Further, the reservoir bed was cored, at 16 sites, using a vibra - corer, to depths of 0.20 to 1.4 m (Photo 4.4). The average specific weight of 5 cores was 1005 kg/m³ and the range was from 911 to 1174 kg/m³. There does not appear to be a systematic variation in the specific weight of these samples, which were taken from the delta and middle third of the reservoir.

Lines of cross - section in the 1977 and 1982 survey were compared with the pre dam topographic survey. The pre dam survey was presented as a topographic map with a five foot contour interval. The total infilling over the period 1932 to 1977 is estimated to be 3,175,764 m³. This represents about 10% of the initial capacity of the reservoir (Hollingshead et al., 1973). The long - term load, given the measured specific weight 1005 kg/m³, is 69,383 t/yr, for the 1932 to 1977 period, or 72540 t/yr for the 1933 to 1977 period. The relevant period of sedimentation is nebulous, because the dam was partially complete and stored much of the 1932 flood wave. Hollingshead's (1969) estimated infilling rate of 75,250 m³/yr, which is equivalent to 75,626 t/yr, is 4 to 8% larger than the more recent estimates. The difference may be due to smaller annual loads since Hollingshead's survey in 1968. However, given the inherent problems in calculating reservoir volumes from a 5 foot contour interval map, the discrepancy must be due, in part, to error. Thus, Hollingshead's (1969) estimated average deposition rate of 75626 t is accepted, because the

Photo 4.3 Glenmore Reservoir delta

Photo 4.4 Obtaining a sediment core from near the delta front

survey methods were more exacting. This value has to be related to sediment input by considering reservoir trap efficiency.

Hollingshead (1969) suggests, on the basis of empirical guidelines from U.S. reservoirs (Brune, 1953), that the reservoir would trap approximately 75 to 95 percent of the sediment inflow. The average load therefore would be in the order of 76850 to 97300 t/yr. However, the reservoirs studied by Brune (1950) have different flow regimes and operating procedures than Glenmore Reservoir. Thus, an alternative approach of estimating trapping efficiency was examined.

The June 1969 flood peak, which is the largest recent flood with available suspended sediment data, was selected for analysis. On June 26 the mean daily reservoir inflow was estimated to be 91.5 m^2/s. The suspended sediment concentration, based on Hollingshead's (1969) rating curve, would be 2067 mg/l. The measured outflow on June 26 had a concentration of 24 mg/l at a discharge of 30 m^3/s, which is about 1.0% of the inflow concentration. Measured concentrations range from 10 to 24 mg/l for discharges from 20 to 30 m^3s during May and June, 1969 (Hollingshead, 1969). There is no relationship between concentration and reservoir inflow or outflow for this data.

Borland's (1971) modified Einstein (1956) equation was used to predict the proportion of the sediment inflow remaining in suspension within the reservoir (Figure 4.9).

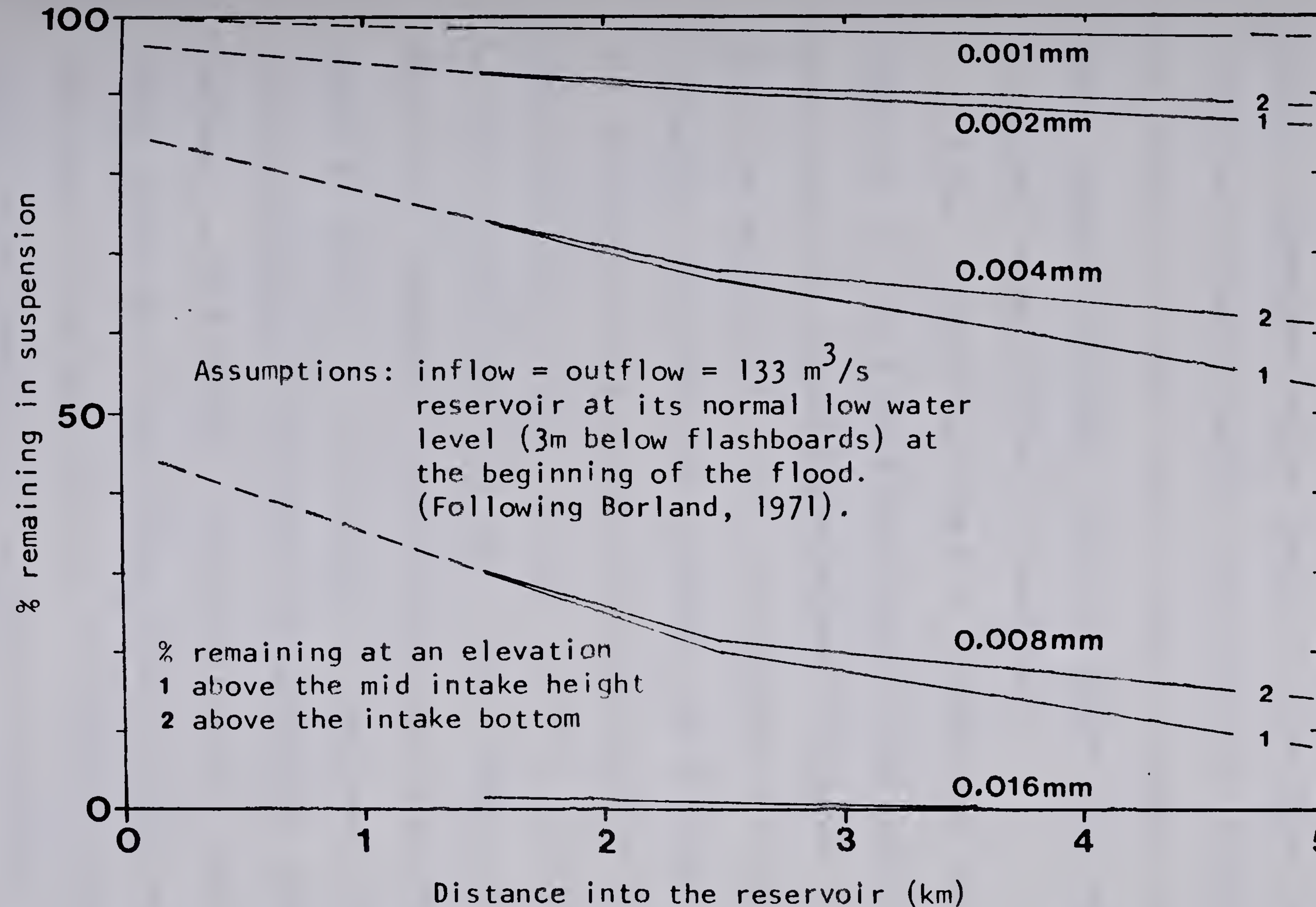


Figure 4.9 Percentage of suspended sediment remaining in suspension within Glenmore Reservoir under flood conditions

The assumption was made that the reservoir was approximately at the normal low water level at the time of the flood. Further, the difference between the normal, pre - flood, water level and the average bed height in given reaches was used to represent water depth in the upper reservoir. In the middle and lower reaches the elevation difference between the reservoir level and the water intakes was used as the depth because sediment falling below this depth was presumed to become unavailable for outflow.

Size - specific rates of deposition were considered in relation to the size distribution character and volume of the inflowing suspended sediment (Figures 4.9 and 4.10). If the fine materials still in suspension were dispersed throughout the water body above the bottom of the intakes, the outflow on the day of the peak inflow would have a concentration of 140 mg/l. Thus, 1,610 t of the sediment would be lost on this day out of a total input of about 22,300 t. Reservoir sediment loss for this extreme event therefore would be about 7% on the day of the peak inflow and outflow. During the short, high discharge period, much of the finer sediment would remain in suspension in the reservoir. Thus, there would be a continuing loss of fine sediment, until these particles settled out. An outflow suspended sediment rating relationship is required to quantify the sediment retention capability of the reservoir any further.

The high trap efficiency of the reservoir is to be expected because of reservoir operating procedures and the

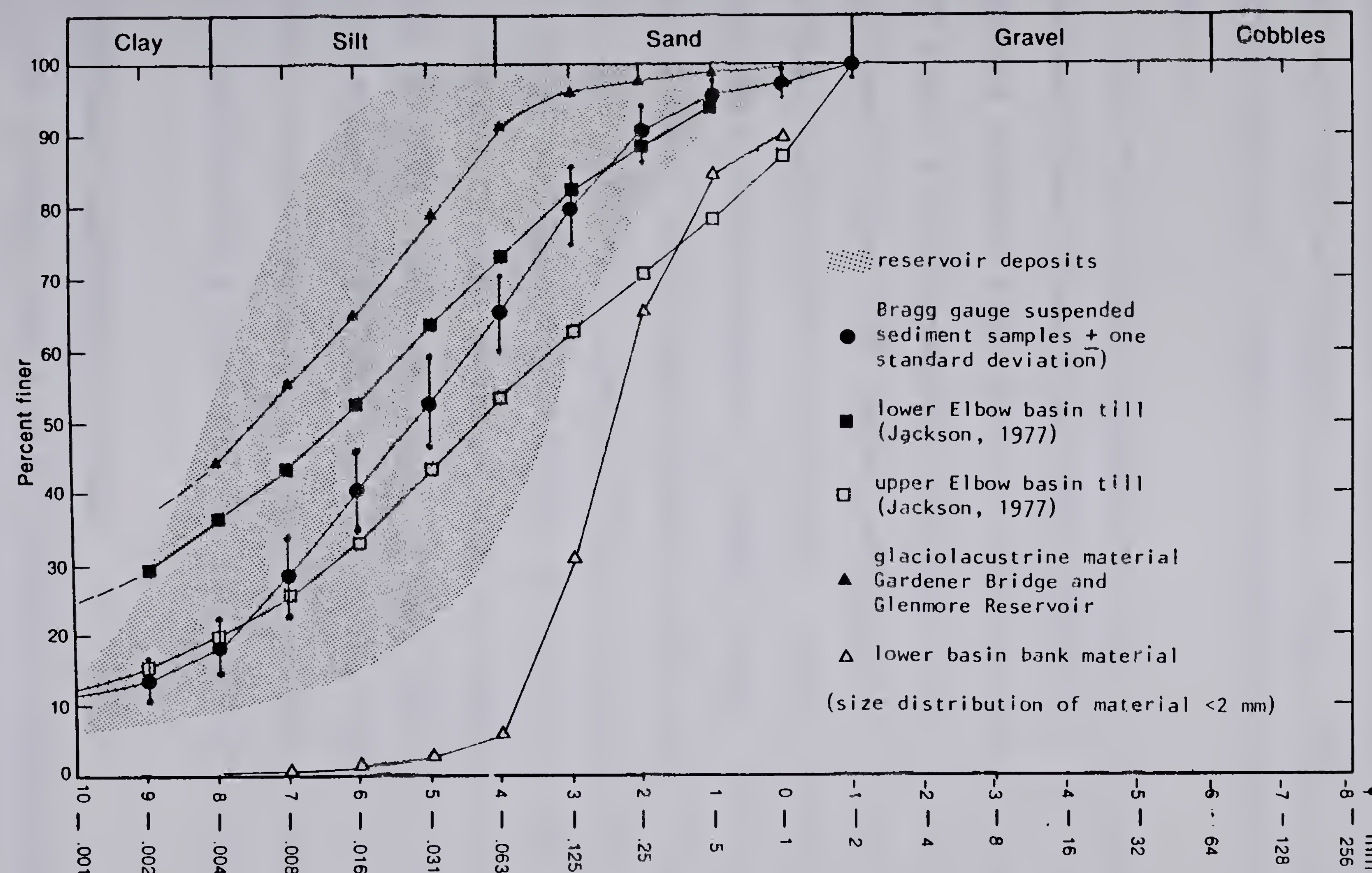


Figure 4.10 Composite particle size distribution curves, Bragg gauge, lower basin river banks and valley walls and Glenmore Reservoir

relative coarseness of the particles. The reservoir is kept about 3 m below full supply level in anticipation of the annual spring flood. The reservoir has not been over - topped since the dam was completed (J. Bouch, Production Engineer, City of Calgary, pers. comm., 1982). The reservoir storage capacity is used to attenuate the downstream flood peak.

The 1969 outflow discharge of 133 m^3/s has been exceeded three times since dam completion. The largest reported outflow (154 m^3/s in 1948) was 15% bigger than the 1969 peak. If the reservoir outflows were of a similar magnitude to the 1969 flood, and reservoir conditions were the same, about 6500 t of sediment would have been lost downstream as the result of these floods. This represents about 0.2% of the total sediment inflow over the period. Thus, the reservoir probably has a trapping efficiency, over the range of flows experienced, of close to 100%.

Given that the reservoir has a trapping efficiency close to 100%, then the long - term suspended sediment load of the Elbow River basin above Glenmore Reservoir is probably about 75600 t/yr, which translates into an upstream unit yield of 62.5 $t/km^2/yr$ for the 1210 km^2 basin.

C. TEMPORAL VARIATIONS IN LOAD

The average load for the period of measurement 1969 to 1979 at Bragg gauge is 18932 t/yr. The estimated long - term average load is 18186 t/yr over the 1935 and 1979 period.

The variation in the annual sediment load is considerably greater than the variation in the annual discharge. The coefficient of variation (standard deviation / mean x 100) of the suspended sediment load is over five times as great as for the annual discharge (load 105%, discharge 19 - 21%) (Table 4.4). The average annual suspended sediment load of the Elbow River at Bragg gauge as calculated by monthly discharges, is log - normally distributed. Return periods and probabilities of exceedence of the annual suspended sediment load have been estimated for both periods (Figure 4.11). The average annual load has a return period of about 3 years (Figure 4.11). Fifty percent of the time the annual load is expected to exceed about 12,000 t. Ninety nine percent of years the load is expected to be less than 100,000 t. The highest measured annual load of 61200 t in 1969, has a 25 year return period.

There is considerable seasonal imbalance in the time distribution of the average monthly sediment load and average monthly discharge regime of the Elbow River at Bragg gauge (Table 4.1, Figure 4.12). Runoff and sediment loads in the winter months represent a small proportion of the average total runoff and sediment budget (13.0% and 0.28% respectively). Both runoff and sediment load tend to decrease from autumn to reach a minimum monthly value in February or March. The bulk of the annual runoff and annual suspended sediment transport occurs in the three month period May, June and July. On average 98.1% of the total

Table 4.4 Estimated annual suspended sediment load and discharge of the Elbow River at Bragg gauge, 1969 - 1979

Period	annual		March to October	
Year	load (tonnes)	discharge (da m^3)	load (tonnes)	discharge (da m^3)
1979	1812	179822	1779	151367
1978	7259	257355	7222	224727
1975	6098	221817	6055	188362
1974	34415	315267	34371	278465
1973	7557	266682	7506	230644
1972	14347	301549	14295	263833
1971	18768	252675	18662	219325
1969	61204	340516	61138	300363
mean	18933	266960	18879	232136
std. dev.	19897	51917	19889	48502
coef. var.	105%	19%	105%	21%

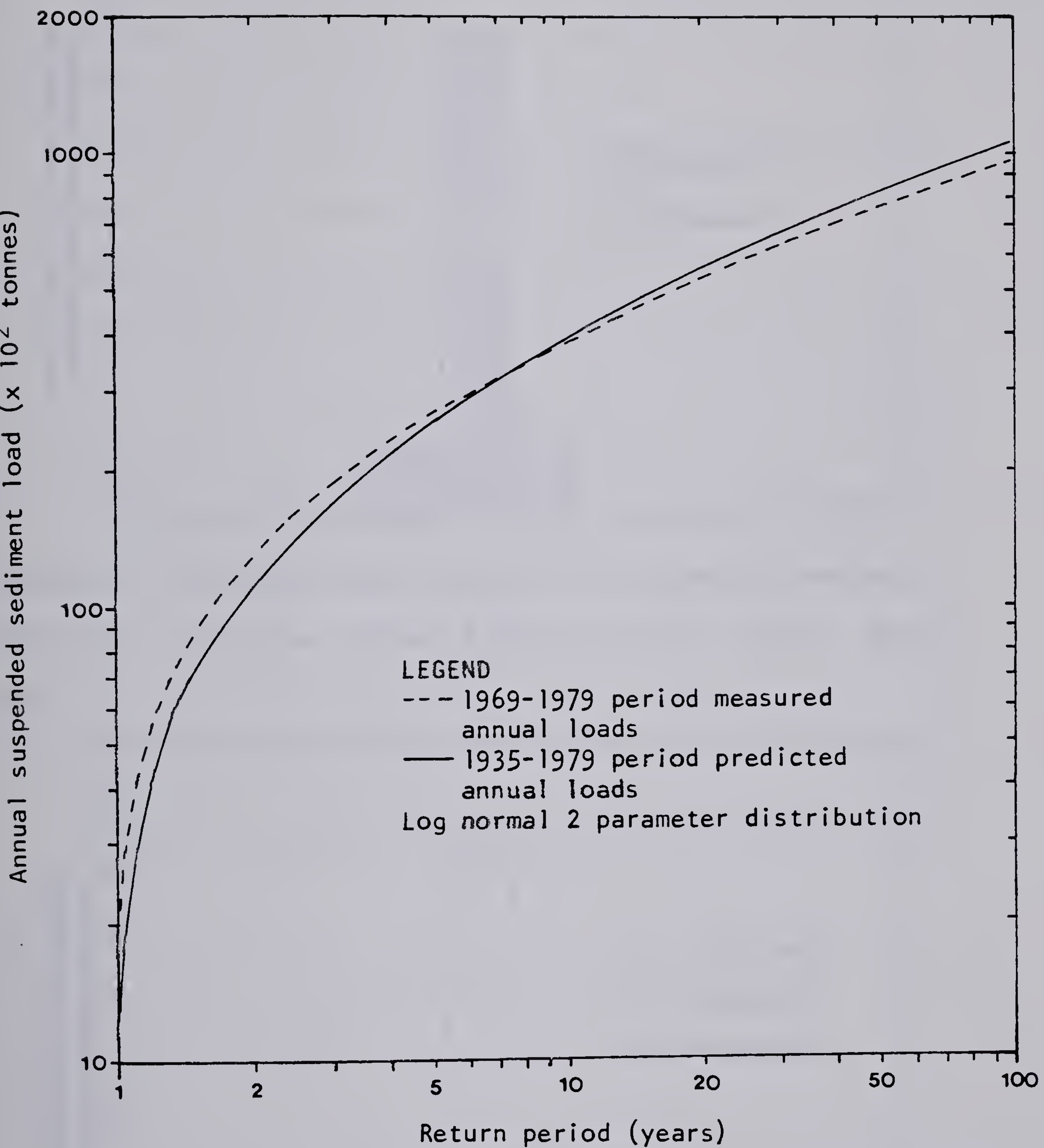


Figure 4.11 Frequency estimates of annual suspended sediment loads, Elbow River at Bragg gauge

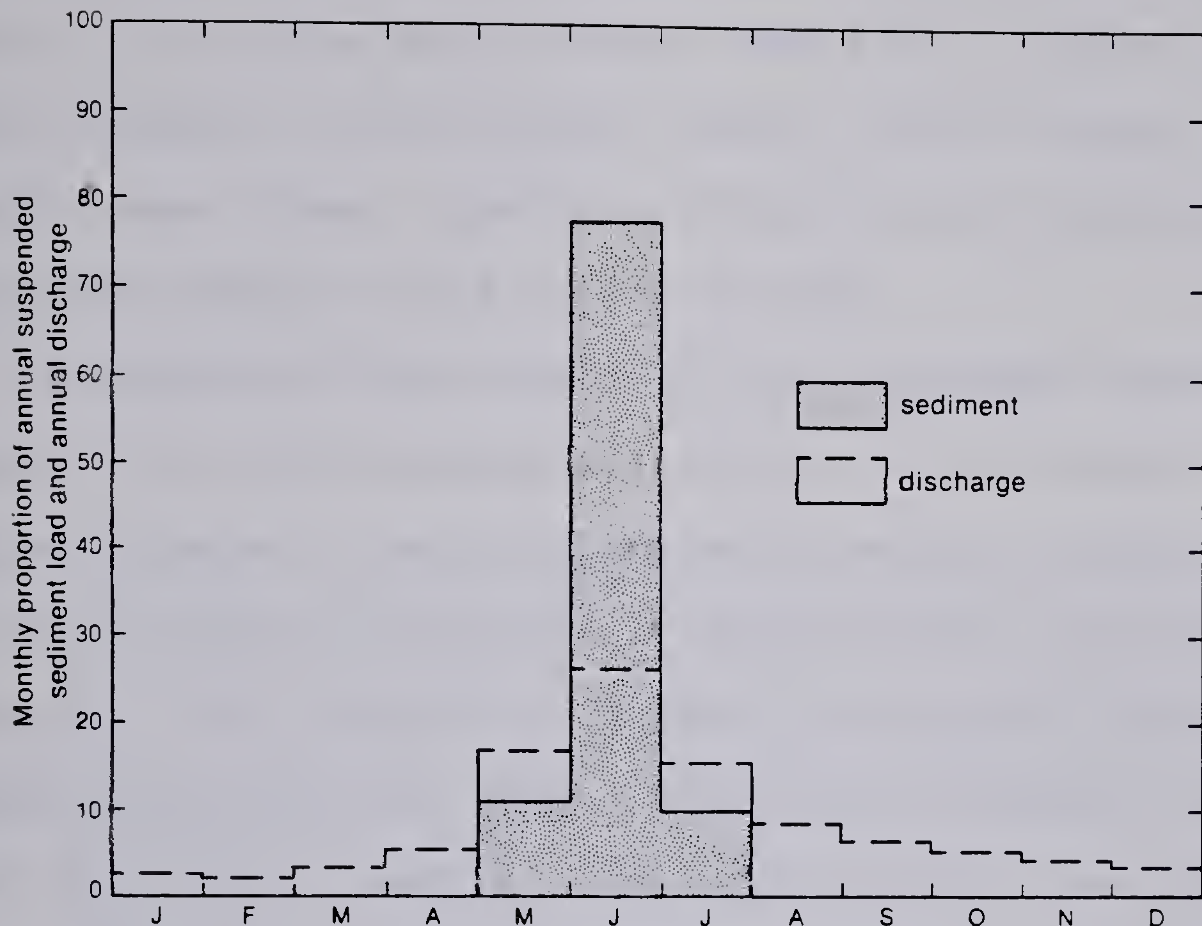


Figure 4.12 Monthly distribution of suspended sediment loads and discharge, Elbow River at Bragg gauge, 1969 to 1979

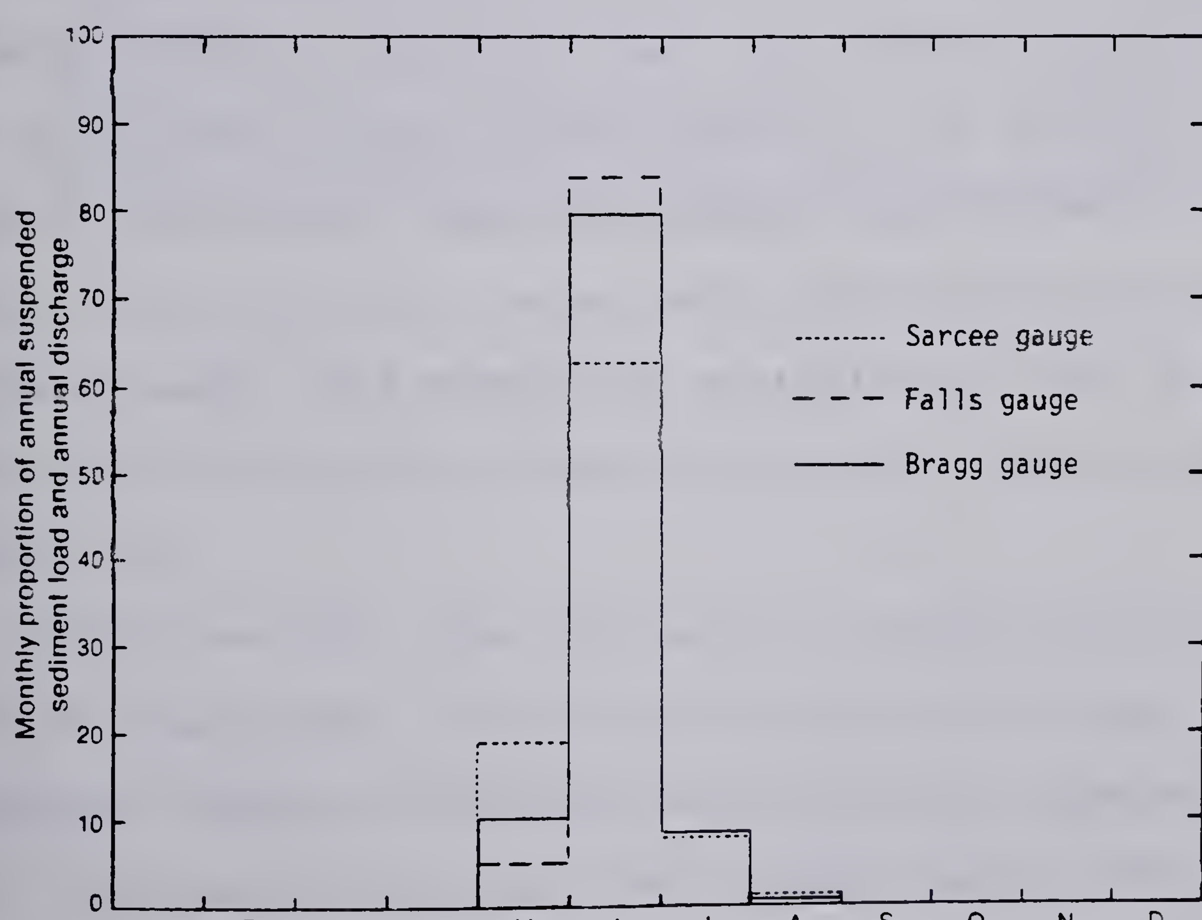


Figure 4.13 Monthly distribution of suspended sediment loads, Elbow River hydrometric stations, 1978 to 1979

annual sediment load and 57.7% of the total annual runoff occur in the three month period (Table 4.1, Figure 4.12). About a quarter of the annual runoff (25.7%) occurs in June whereas over three - quarters (77.8%) of the average annual suspended sediment load occurs in June.

The monthly distribution of the suspended sediment loads of the other gauges are similar to the Bragg Creek regime. However, the distribution of monthly loads appears to flatten going downstream to Sarcee gauge from Falls gauge (Figure 4.13). Further, individual tributaries may behave somewhat differently. Gauge, McLean and Silvester creeks have the highest recorded suspended sediment concentrations in the Elbow River basin. Concentrations in the order of 1500 mg/l were measured in these streams following convectional rainstorms (Photo 4.5). These storms are usually of short duration, thus the sediment yield is small for individual storms (in the order of 0.2 to 2.0 t). However, the large number of convectional storms cumulatively produce a relatively large monthly suspended sediment load. This results in a relatively more balanced distribution of monthly loads through the spring and summer (Table 4.3).

Daily records, from the March to October period 1969 to 1979 at Bragg gauge, show that the high discharges, high suspended sediment concentrations and high suspended sediment loads generally occur over a short period. For 99% of the time the concentration was less than 300 mg/l, for 95%

Photo 4.5 McLean Creek following a convectional shower, July, 1979

of the time concentration was less than 71 mg/l and 50% of the time concentration was less than 3 mg/l (Figure 4.14).

There are usually a few, relatively minor, increases in discharge and concentration in the early spring which precede the main spring flood. There is usually a rapid rise in mean daily discharge with high suspended sediment concentrations. Maximum concentrations occur at the time of the annual peak discharge or on the previous day (Figure 4.15). Following peak discharge, concentration decreases dramatically. The suspended sediment concentration and loads tend to be far greater on the rising limb of the hydrograph than following the peak discharge (Figure 4.16). As a result a considerable proportion of the annual suspended sediment load is transported in a few days.

The cumulative proportion of total annual load transported with time increases at a declining rate so that at Bragg gauge 50% of the average annual load occurs in 3.6 days and 72% occurs within ten consecutive days. At Sarcee gauge the peak discharge period produces a relatively smaller proportion of the total annual load than at the upstream stations (Figure 4.17).

D. DISCUSSION

Three related facets of the suspended sediment regime are the focus of discussion: (a) the sources of sediment and (b) the controlling processes which determine the character of the suspended sediment regime and (c), comparative

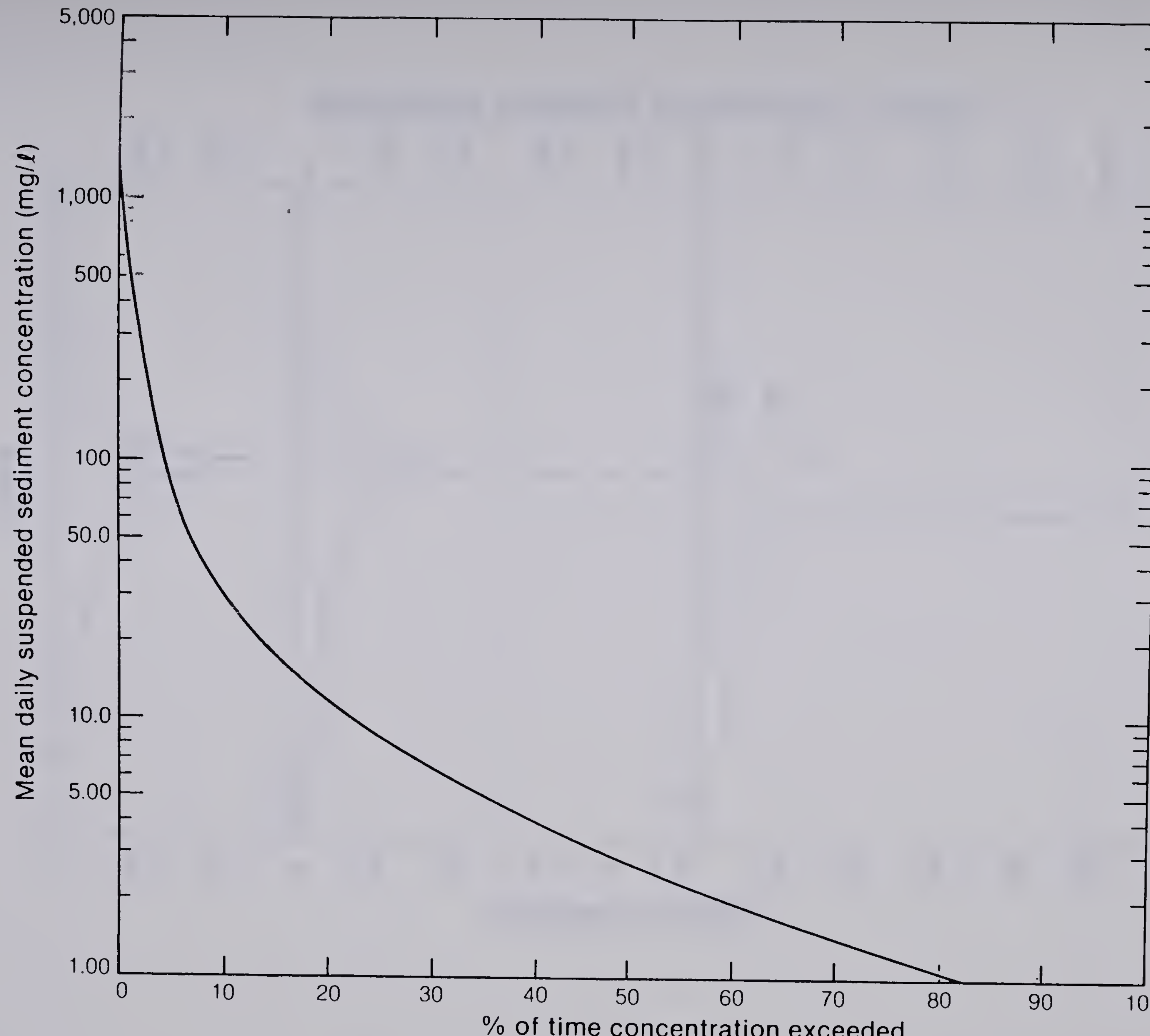


Figure 4.14 Concentration duration curve, Elbow River at Bragg gauge, March to October, 1969 - 1979

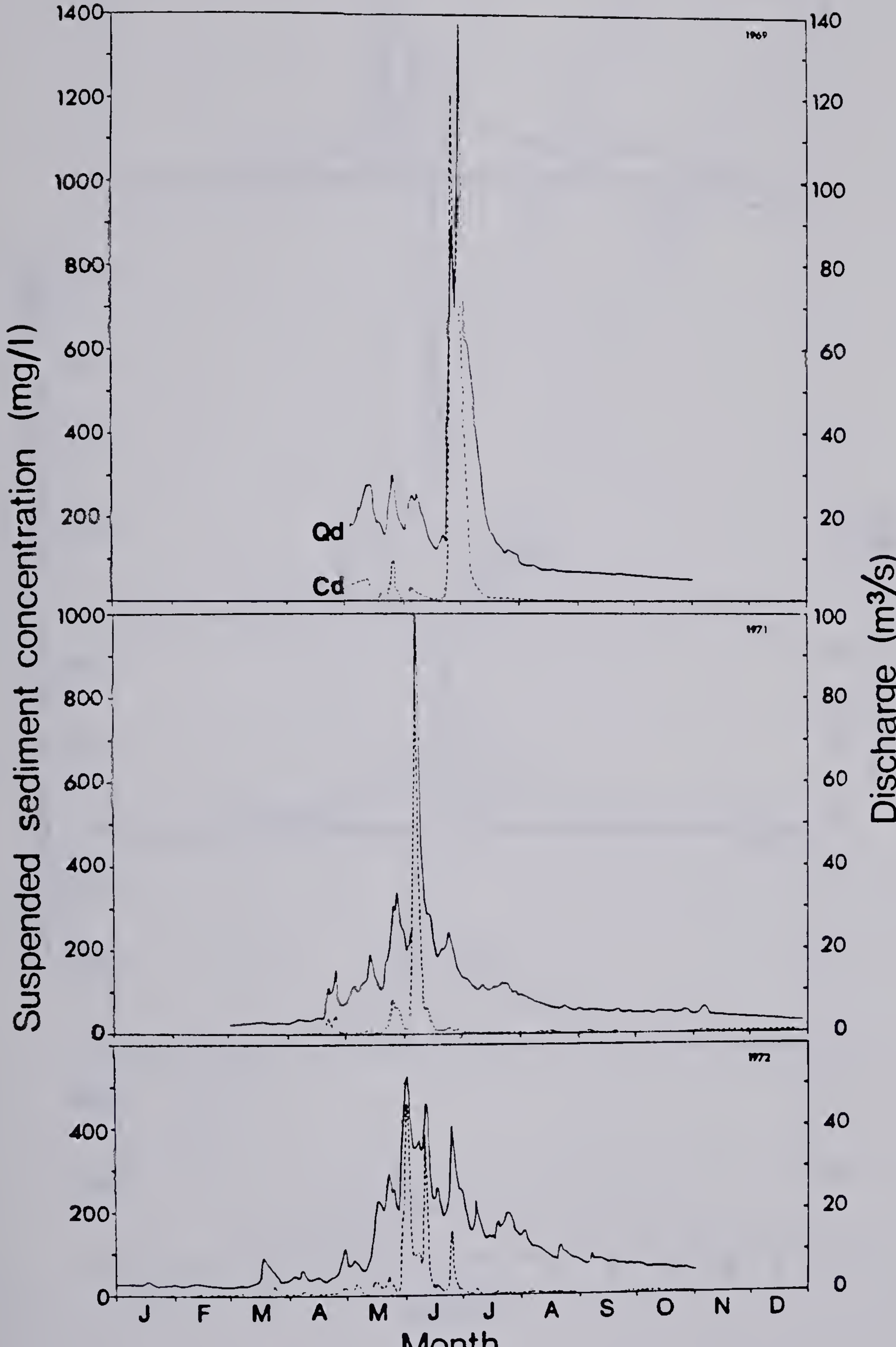


Figure 4.15 Mean daily discharge and suspended sediment concentrations, Elbow River at Bragg gauge, 1969 - 1979

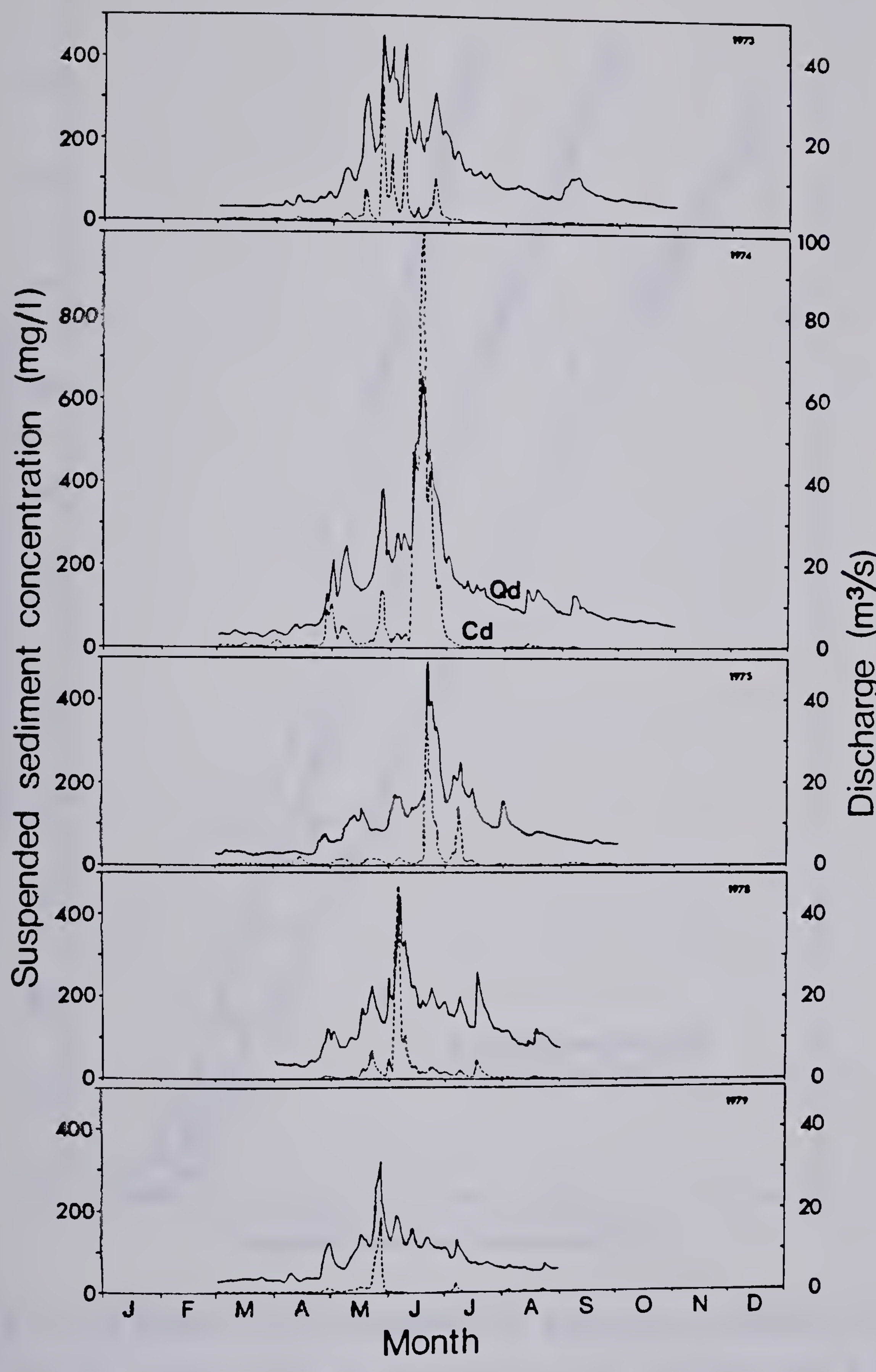


Figure 4.15 Continued...

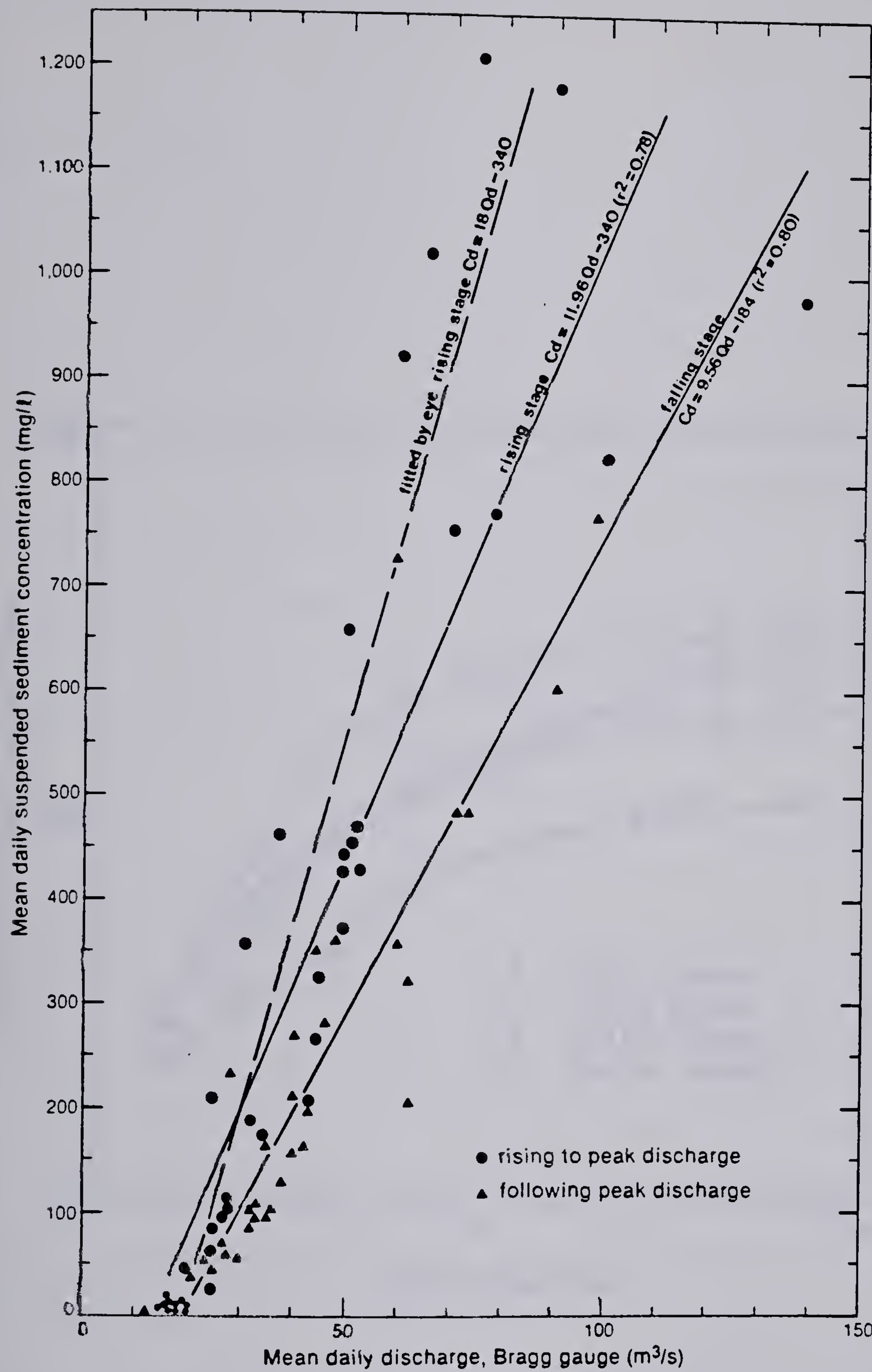


Figure 4.16 Mean daily suspended sediment concentration against mean daily discharge for a five day period before and after maximum daily discharge, Elbow River at Bragg gauge, 1969 - 1979

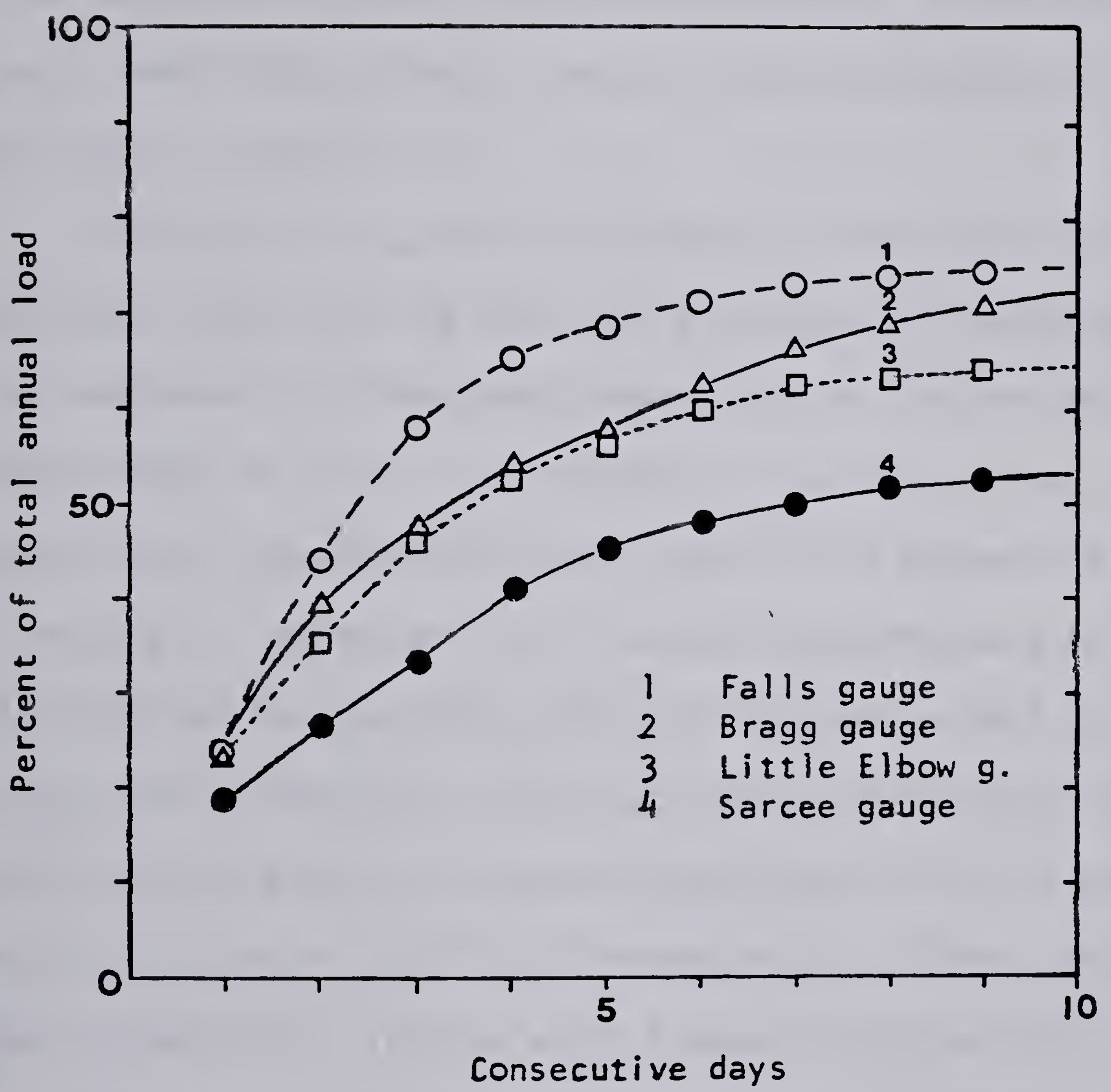


Figure 4.17 Cumulative percentage of the total annual suspended sediment load transported over time at Bragg gauge

magnitudes and frequencies of suspended sediment transport.

Sediment sources

An interpretation and mapping scheme in which the type of erosion process, relative rate of action, size characteristics of the erosion product and relative rate of delivery to the stream network, is determined for individual erosion sitess, over large areas, using aerial photograph, has been described in Chapter 2.

A detailed air photo analysis of the Elbow River basin, undertaken previous to the field program, suggested that the major sediment sources and processes in the mountains are surface wash of non - vegetated colluvial slopes, plus channel and riparian erosion. Loads were expected to be low, but variable, because: (a) the wash processes are effectively limited to barren colluvial slopes with a direct link to the stream network; (b) mechanical weathering of limestone and dolomite bedrock produces limited amounts of fine clastic material; (c) the mountain river channels appear relatively stable with coarse textured bed material; (d) impingement of the river upon fine - matrix morainic deposits are infrequent; (e) mass movements infrequently reach the channel network; and (f) sediment sinks may form upstream of debris jams in some of the major tributaries.

The field observations largely substantiate the conclusions of the air photo interpretation. The long - term basin yield of the mountains is about 11 $t/km^2/yr$. Between

75 and 90% of the suspended sediment load of the Little Elbow River basin was derived from tributary sources. The remainder of the load was derived from erosion of a morainic deposit which slumped into the river channel, as the result of saturation during spring melt (Photo 4.6). Detailed bank profiling (Hudson, 1982a), which was used to estimate the volume and type of bank material removed by erosion, and cross sectional surveys, indicated that only local bed and bank erosion occurred in 1978 and 1979. A number of riparian, gullied, morainic deposits, which give the appearance of being major sediment suppliers, did not produce surface runoff and were protected from river erosion by bouldery lag deposits along the water's edge (Photo 4.7). Other cliffs were similarly protected.

The major variations in suspended sediment yields in the mountains of the Elbow River basin are explicable in terms of the sediment sources and sinks identified in the air photo analyses and field program. Annual load and drainage area are strongly related statistically (Figure 4.7):

$$\text{Load (t)} = 7.97 \text{ area } (\text{km}^2)^{1.07} \quad (r^2=0.94) \qquad ...(4.1)$$

The relationship between load and area should be used cautiously. The two mountain stations, the upper Elbow River and the upper Little Elbow River, which have greater loads than predicted, both have significant point - source contributions. However, basins which plot on, or below, the "best fit" line have variable characters. Basins WG2 and WG4

Photo 4.6 Slumped till deposit, Little Elbow River

Photo 4.7 Bouldery lag deposits and vegetation may protect valley wall deposits

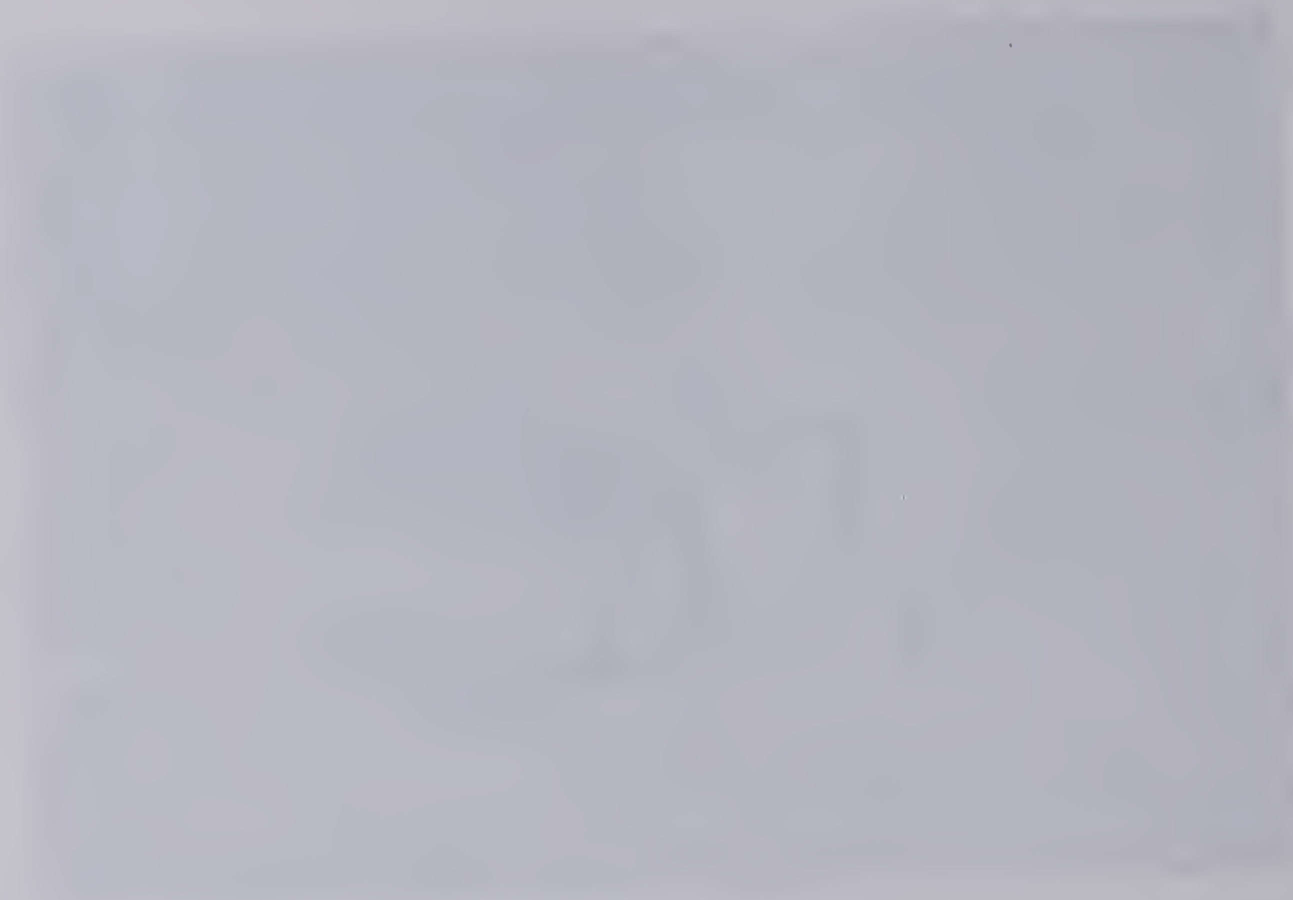
are both barren, calcareous, colluvium tributaries, and Set Creek has significant point sources but has a number of sediment traps. The Little Elbow River has significant point sources (10 to 25% of the annual load) as well as large non - contributing areas (at least during the period of observation in 1978 and 1979) such as Shoulder Creek (23.1 km^2) and other unnamed tributaries.

Air photo analysis suggested that the major suspended sediment sources in the Falls gauge sub - basin (Figure 4.1) are channel and riparian erosion. Sheet wash processes may occur in the barren, colluvium, mountain portion of the sub - basin. However, although the upper foothills include areas of fine grained colluvium derived from fine clastic bedrock, surface erosion processes are limited by the dense vegetation cover for almost all of the remainder of the basin. The localized exceptions of surface erosion are small areas of denuded mass movement scars. In addition, roads and trails, which traverse easily erodible shales and surficial materials, may provide sediment to the stream network (Photo 4.8). Mass movements of the upland slopes generally do not reach the stream network, with a few notable exceptions.

The suspended sediment load of the Falls gauge sub - basin was determined by measurement of the sub - basin input and output and by measurement of the load of two tributaries within the sub - basin (Table 4.3; Figure 4.1). The loads of unmeasured tributaries were predicted by the previously described load - area model. Measured loads at Falls gauge

Photo 4.8 A motor cycle trail crossing McLean Creek

Photo 4.9 Sediment from erosion of road margins in Gauge Creek directly enters the Elbow River above Falls gauge



and from Ford and Gauge Creek are reasonably described by the model when the annual values are transformed into a long - term estimated load (Figure 4.7). It was thought that the similarity in characteristics of the mountain basins would make the model valid for those stations. However, the load from south Glasgow Creek may be under - estimated because of channel erosion of till in this valley. Also, the load from Nihahi Creek may be over - estimated because the stream only had a trace of flow during the period of observation in 1978 and 1979.

The long - term average suspended sediment load of the Falls gauge sub - basin, which represents the upper foothills, is 16.3 t/km²/yr. In the order of half (45 to 60%) of the annual load of about 4000 t is derived fom tributary streams. The remainder of the load (about 2050 t) is derived from bank erosion from the 27.6 km reach between sampling stations (Figure 4.1). Tributary stream loads also were almost exclusively derived from channel or riparian erosion. Erosion of road margins and trails appears to be of local importance. Convectional storms resulted in the erosion of road cuts in shale bedrock and silty sandy till in upper Ford Creek and in the middle of Gauge Creek. The Ford Creek load is, however, largely trapped by numerous beaver dams and debris jams where considerable siltation occurs (Photo 2.14). The load from road margins in Gauge Creek is directly routed to the Elbow River (Photo 4.9).

The Bragg gauge sub - basin has a variety of tributary characteristics. Two of the major tributaries, Canyon Creek and Prairie Creek, drain from the mountains and have tributary streams in the foothills which drain the Paleozoic bedrock outliers, Moose and Prairie mountains. Air photo analysis suggested that surface wash, of the barren colluvium slopes, may be important. However, because of the generally dense vegetation cover, the dominant sediment source was thought to be channel and riparian erosion of the extensive fine textured colluvial deposits and fine textured till which cover the area, and erosion of the glaciolacustrine material in the lower reaches above Bragg gauge. The air photo analysis suggested that the roads and trails in the McLean Creek and Sylvester Creek basins could be significant sediment sources.

The major tributaries of the Bragg gauge sub - basin were sampled in detail in May and June 1978 and 1979 and miscellaneously for the remainder of the study period. Annual and long - term loads were estimated from the May and June data by comparison with the Bragg gauge May and June loads in relation to the annual long - term load (Table 4.3). Mclean Creek was intensively sampled during snow melt and rain storms throughout the April to August period. Three other significant tributaries were not sampled, either because of access difficulties (Iron and Silvester Creek) or because little runoff was observed((Powderface Creek) (Table 4.3). The suspended sediment load of the larger

unmeasured tributaries was estimated by area - weighting with similar measured tributaries (Table 4.3). The remaining 53 km^2 (or 15%) of the Bragg gauge sub - basin consists of very small tributaries, valley side slopes, low terraces and floodplain.

About 80 to 90% of the suspended sediment load of the Bragg gauge sub - basin is derived from areas other than the major tributaries (Table 4.3). Undoubtedly, some suspended sediment is derived from very small unmeasured tributary streams, such as Connop Creek (3.8 km^2). However, the loads from such streams are thought to be insignificant. Further, much of the remainder of the sub - basin is separated from the river channel by broad alluvial plains or terraces. Thus, they also cannot contribute to the suspended sediment load of the river at the observed level of flooding. It is therefore concluded that the bulk of the suspended sediment load from the Bragg gauge sub - basin is derived from channel and near - channel sources.

There is considerable spatial variation in the tributary suspended sediment loads. Canyon Creek, which originates in the mountains and flows between the Moose and Prairie Mountain outliers, has a very low suspended sediment load and yield (Table 4.3). McLean Creek, on the other hand, has about 10 to 20 times the long - term yield of Canyon Creek (7.7 to 22.8 $t/km^2/yr$, Table 4.3). The yields of other tributaries lie between these extremes. Canyon Creek has a low suspended sediment yield for two main reasons. The

sediment supply is limited (carbonate colluvial parent material and vegetation protection of fine textured lower slope deposits), and stream flow is ephemeral (because of seepage losses). McLean Creek flows through an area of highly erodible surficial materials (Figure 2.7). These materials are frequently exposed to surface erosion in well used seismic trails which cross the stream network. Roads and other trails (motor cycle and livestock) also cross the stream at numerous points. In addition to this accelerated erosion, the stream cuts into the surficial material units and recessive bedrock.

Over the long - term the load at Sarcee is about 75600 t/yr compared with 18200 t/yr at Bragg gauge. The net load of the Sarcee gauge sub - basin of 57400 t is largely derived from river bank and riparian erosion. The lower basin tributaries were not gauged during the course of the study. However, miscellaneous observations suggest that the tributaries contributed very limited runoff and sediment to the Elbow River system in 1978 and 1979. However, this may not be the case over the long - term since larger floods, with a major rainfall component, produce significant runoff from the lower basin (Chapter 3 and Appendix 2). It is hypothesised, therefore, that for small floods the sub - basin suspended sediment load is derived almost exclusively from near the river channel. As the magnitude of the flood discharges increase the upland component probably increases. Further, there may be a significant snowmelt contribution to

runoff in spring, well before the peak discharge at Bragg gauge, which may move significant suspended sediment load to the Elbow River.

An initial premise of this thesis was that surficial materials in the lower Elbow River basin were derived from Laurentide glaciations rather than from Cordilleran glaciations. Hence, the partly Canadian Shield - derived lower basin surficial materials would be clearly distinguishable from the local, largely calcareous, upstream deposits. However, recent reports by Jackson (1977 and 1980) show that the boundary between the Laurentide and Cordilleran deposits is not near the foothills - plains boundary, as suggested by Seagel (1971), but much further east, near the lower boundary of the Elbow River basin. As a result, it was not possible to use petrology to differentiate sediment sources.

Some indication of possible sediment sources may be obtained from a comparison of reservoir deposits and the particle size distribution of upstream sediment sources (Figure 4.10). The average particle size composition of the reservoir deposits has not been fully defined. Hollingshead (1969) stated that the average composition of the reservoir deposits is the same as the composition of suspended sediment at Bragg gauge: approximately 15% clay, 50% silt and 35% sand. He also noted minor gravel deposits in the upper delta. However, his samples were taken from the reservoir delta, and did not include the lower reservoir, where over

half of the deposition has taken place. A core sample from the delta front had an average composition of 8.8% sand, 81% silt and 11.1% clay. Further downstream the values were 0.25% sand, 47.4% silt and 52.4% clay. If these samples represent the particle size distribution of the deposits in the respective areas, then the average composition of the reservoir sediments can be estimated from the amount of deposition in each zone, using a dilution analysis. The average composition of the reservoir deposits are thus estimated to be approximately 23% clay, 56% silt and 21% sand. Because the reservoir deposits appear to be generally finer than the inflowing sediment from upstream at Bragg gauge, it is concluded that the reservoir materials are largely derived from the fine textured valley wall deposits. In addition, coarser sediments are present. These particles may be derived from the valley wall deposits as well as from river bank erosion.

Schumm (1977) suggested that drainage basins may be divided into three zones. Zone 1, the upper basin, is the primary zone of runoff and sediment production. Zone 2, the mid basin, is primarily a zone of fluvial sediment transport. Zone 3 is a sediment sink. In terms of runoff and sediment production, the Elbow River basin tends to conform to this idealized fluvial system. Theoretically, upland erosion processes determine the sediment yield in zone 1 (Hadley and Schumm, 1961). This is largely the case in the mountains of the Elbow River basin. Further, the major

sediment sources in zone 2 are channel and riparian erosion, which is the case in the foothills and plains of the Elbow River basin. Glenmore Reservoir obviously represents a sediment sink. This downstream variation in sediment sources may well be a general Eastern Slopes phenomenon. Luk (1975) concluded that three sediment sources appear to be significant contributors in the 19710 km^2 Bow River basin above Bassano Dam: (a) excessively erodible glacial / alluvial deposits and bedrock along the major drainage channels; (b) cultivated and excessively grazed areas in the foothills; and (c) steeply sloping coulee sides, ditches, and cultivated land on moderate slopes in the prairies. Neill and Mollard (1982) undertook a reconnaissance survey of erosional processes and sediment yields in the upper Oldman River basin, in south western Alberta. The 4000 km^2 basin is composed of mountains, foothills and plains, which are similar physiographically to the Elbow River basin. The suspended sediment yield of the Oldman River basin is also similar (70 versus 62.5 $t/km^2/yr$). Their study concluded that the greater part of the suspended sediment load in the Oldman River was probably derived from stream bank erosion and landslides into the river, and some tributaries, of the lower foothills and plains of the basin. Channel instability in the mountains was thought to supply mainly bedload material. Gullying, which was of secondary importance, was often associated with disturbed areas, particularly in the finer textured surficial material areas of the lower Oldman

River basin.

Observations in other areas suggest that the predominance of channel and riparian erosion is to be expected when the drainage basin has a protective vegetative cover, as in the mountain valleys and foothills of the Elbow River basin, or when a combination of low slopes and vegetation inhibit erosion and transport, such as in the plains area of the basin (e.g. Coldwell, 1957; Glymph, 1957; Langbein and Schumm, 1958; Hadley and Schumm, 1961; Swanston and Dyrness, 1973; and Fredricksen, 1970).

Controlling processes

The short duration of high suspended sediment concentrations, resulting in the bulk of the annual load being transported in a few days, and the generally greater concentrations and loads on the rising limb of the hydrograph rather than following the peak discharge, are features which are common to a number of areas (e.g. Colby, 1956; Guy, 1964; Porterfield, 1972; and Walling, 1977b). These features can be explained by the controls imposed by the hydrologic regime and the sediment supply regime in the Elbow River basin.

In the mountains of the Elbow River basin, the suspended sediment regime is largely determined by surface wash processes of barren, colluvial, drainage basins (Photos 2.3, 2.15 and 2.16). Runoff from these basins is largely generated by snow melt, which occurs over a period of days

to weeks. These smaller streams are ephemeral. In addition, point - sources, such as debris avalanches and slumps of the valley walls, tend to occur during the spring. These deposits may move downslope so as to constrict the stream channel (Photo 4.6). Surface wash and additional minor slumping may occur during this period. The following significant discharges begin to remove the saturated deposits, which are relatively easily erodible. With time the constriction is opened, stream competence decreases and sediment supply decreases. Thus, the bulk of the transport occurs in a short period in spring.

Nanson (1972) described an identical sequence of events in Bridge Creek and Two O'Clock Creek, in the mountains of the Eastern Slopes in the upper North Saskatchewan River basin, Alberta (Figure 1.1). High banks of glacial debris failed in early spring, as the result of saturation. Surface wash processes followed failure. The mass movement deposits were removed from the stream channel by the rising spring flood. As a result, the transport of sediment for a given discharge was less after the seasonal flood peak than previous to the peak.

The importance of tributary and main river channel erosion and riparian erosion increases downstream. In early spring, discharge may increase without producing a significant increase in sediment concentration. At the time of the main spring flood, which rises rapidly, suspended sediment concentrations increase very rapidly to produce a

peak concentration at or before the time of the peak discharge. Concentration decreases rapidly following the peak discharge and, typically, concentrations on the falling limb are far less than during the rise to peak, for given discharges. These feature are commonly explained by surface erosion processes and the timing of tributary inflows (e.g. Colby, 1956; Guy, 1964; and Porterfield, 1972). However, in the Elbow River basin upland erosion sources provide only a small proportion of the suspended sediment yield. The major sediment sources are river bank and riparian erosion. Carson et al., (1975) explained how these types of rating relationship features are a product of river bank erosion in the dominantly forest covered, Eaton River basin, Appalachian Mountains, Quebec. The time variations in the concentration - discharge relationship can also be explained by bank and riparian erosion processes in the Elbow River basin (Hudson, 1982a and 1982b). Several stages have been recognised (Figure 4.18).

In the early spring period, when stream discharge increases one - or two - fold from the winter low flow period, the river banks and bars typically are sheltered by stranded ice which forms and degenerates in place (Photo 4.10). Cantilevering of the material from the banks, and ice - pull of bank materials occur, which provide relatively small additions to the toe accumulation (Photo 4.11). The ice melts out almost completely before the first substantial rise in discharge. Minor load and discharge peaks occur as

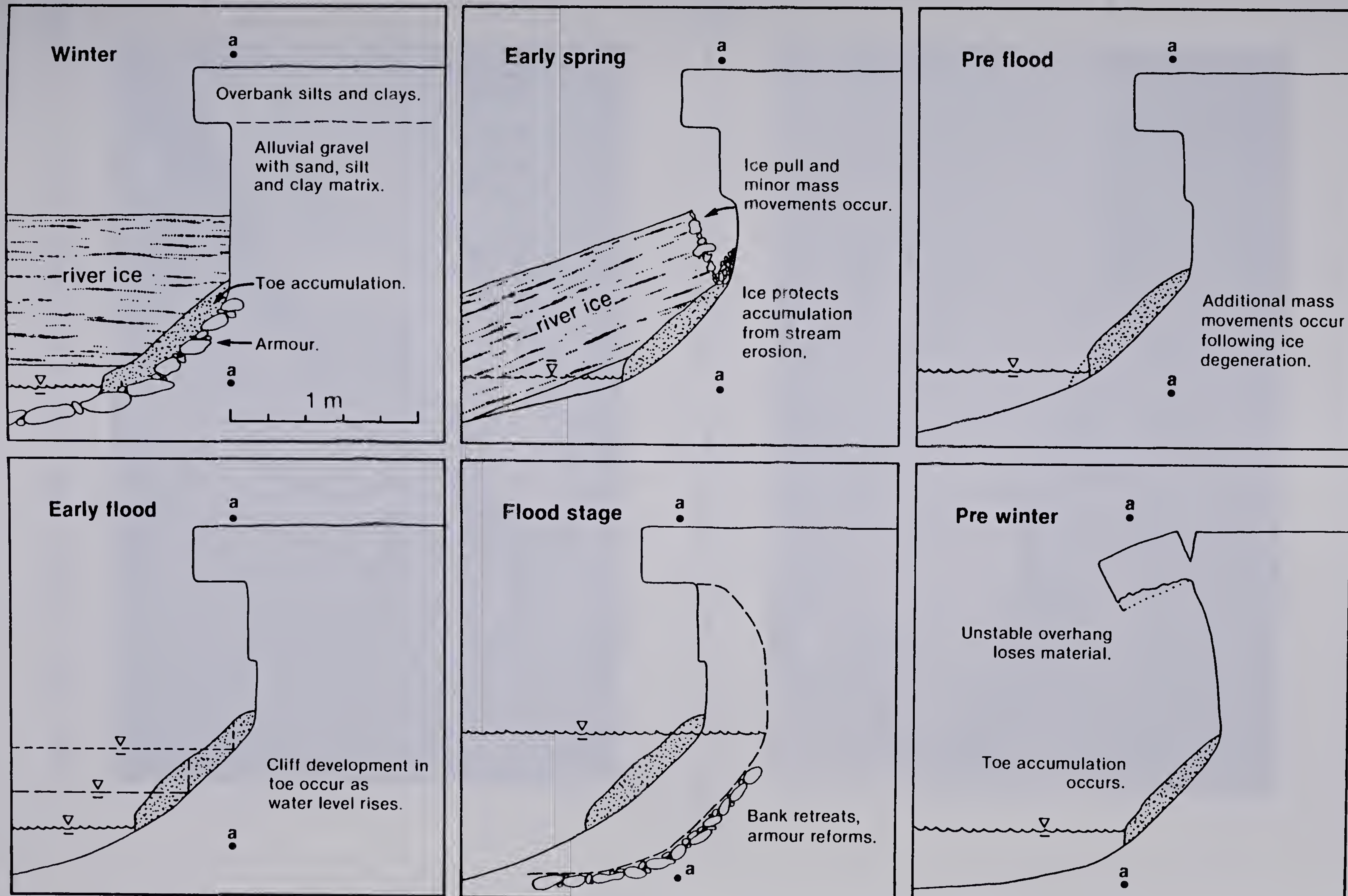


Figure 4.18 A conceptual bank erosion model

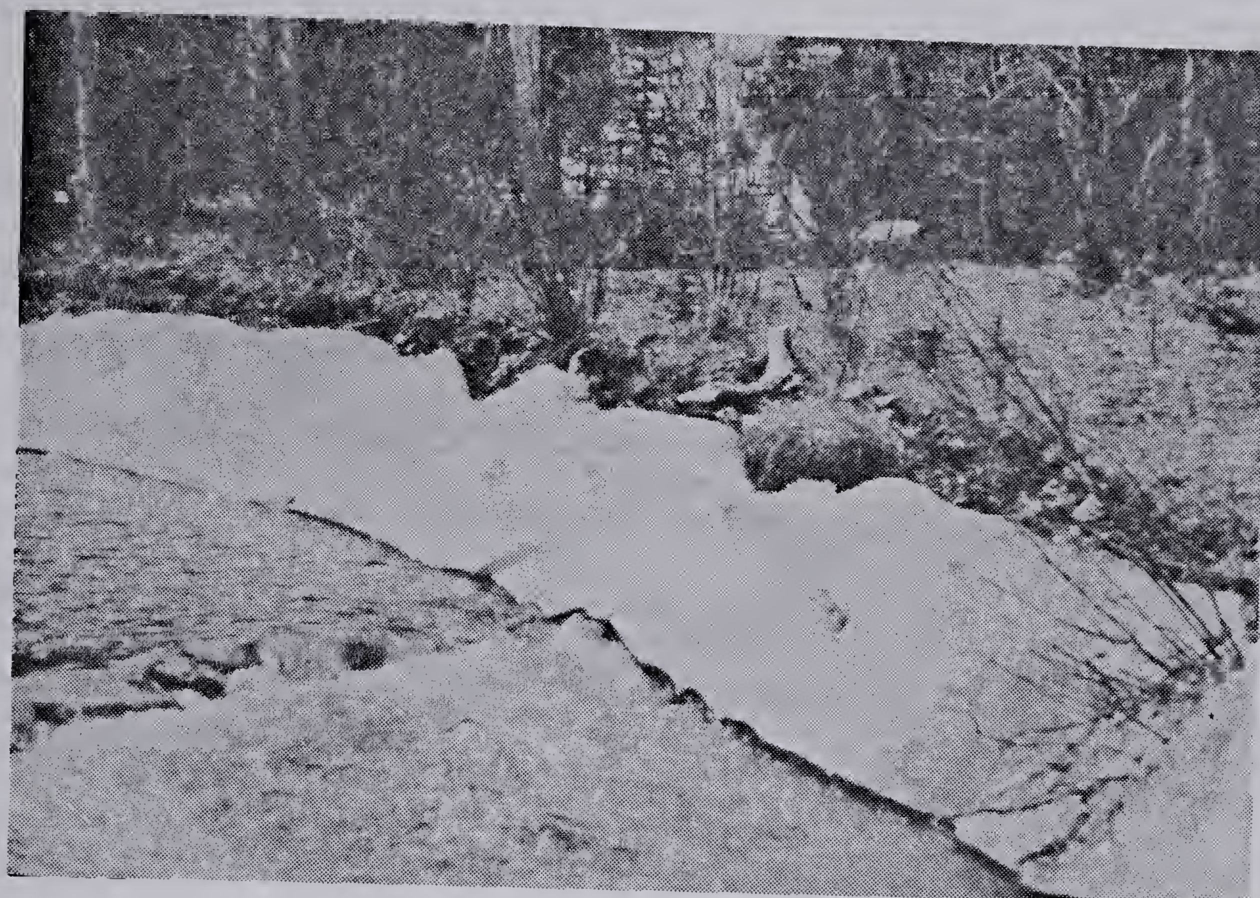


Photo 4.10 Ice may protect river banks from stream erosion

Photo 4.11 Cantilevering of ice may remove river bank material

the result of tributary inputs and minor flushing of the toe accumulations (Photo 4.12).

The main spring flood is, to varying degrees, a basin - wide phenomenon. Greater discharge results in a rise in stage so that weathering products are removed from the toe of banks and cliffs. Removal takes place progressively, producing small cliffs in the toe deposits (Photo 4.13). At high stages, which correspond to discharges of about 20 to 30 m^3/s (or a 1.05 to 1.25 year return period flood), the vertical face of the bank is attacked by the stream. Erosion of the vertical face may be initially easier, because of weakening of the bank by weathering. This hypothesis is based on upland erosion observations (Emmett, 1970). The removal of all or part of the easily erodible surface layer of bank material may take place at any time during high flows. A coarse pavement is developed, or exposed, at the base of the river banks during high flows (Photo 4.14). This pavement may limit bank erosion. Also, the undermining and collapse of vegetation mats may protect the river bank (Photo 4.15).

Following the peak discharge the suspended sediment concentration, hence load, decreases dramatically. This phenomomen is attributed to a decrease in sediment supply. This decrease in supply is exemplified by the case of twin peaks in discharge over a short period producing considerably less load on the second peak (Figure 4.15). Carson et al., (1973) and Hooke (1979) also noted a similar

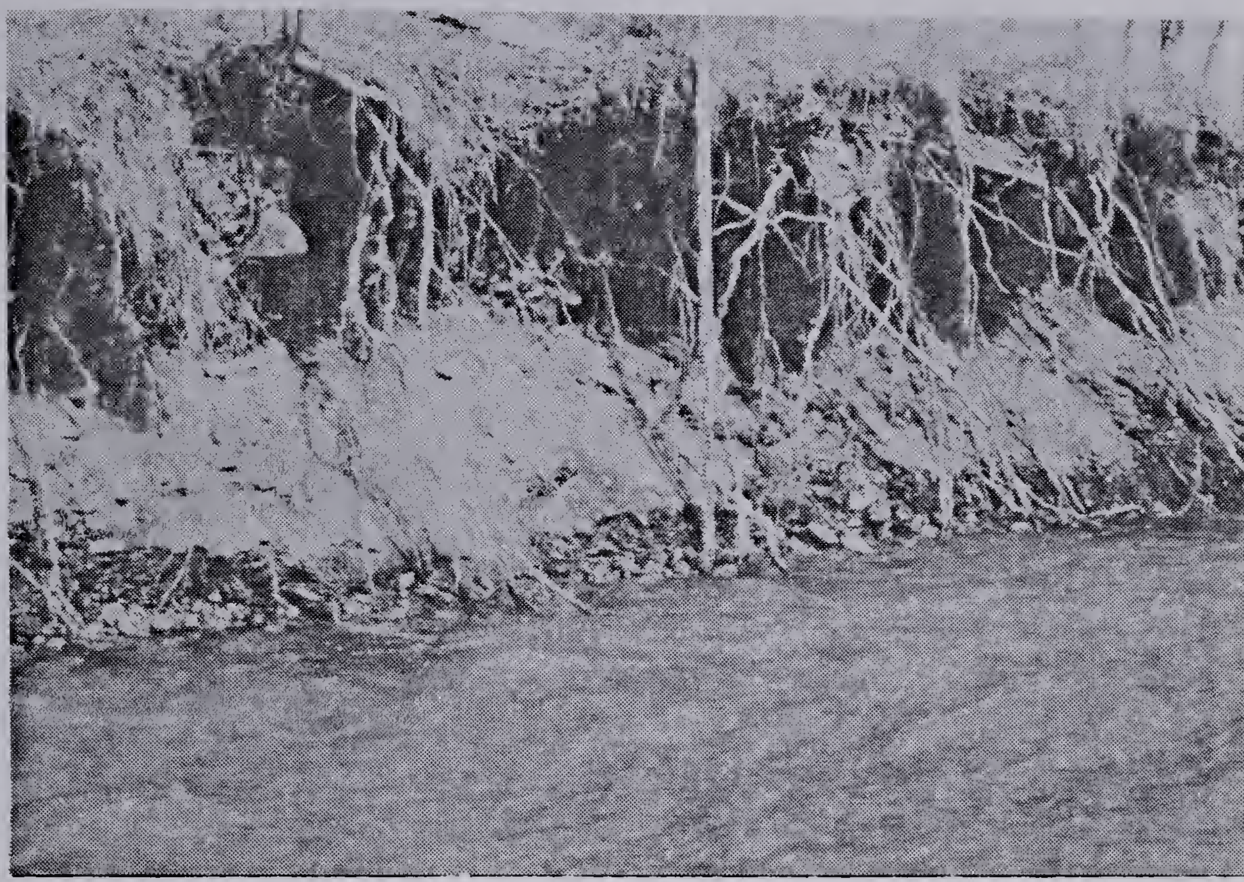


Photo 4.12 Minor flushing of bank toe accumulations occur as the water level rises

Photo 4.13 Major flushing of bank toe accumulations occur at higher water levels

Photo 4.14 A coarse pavement developes, or is exposed, during high flows

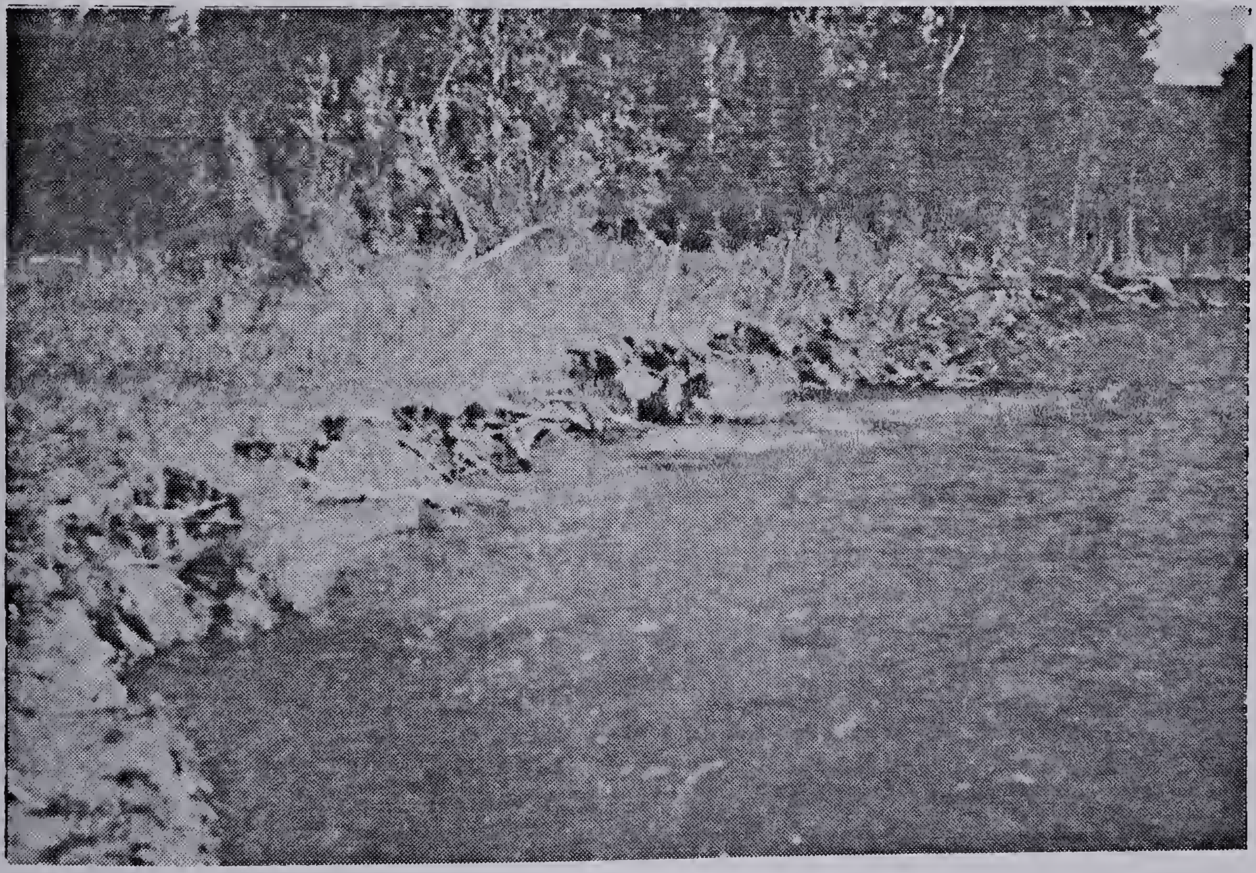


Photo 4.15 Bank erosion may be inhibited by undermined vegetation mats

sediment supply constraint in their rating relationships.

The next stage of the bank erosion cycle is the post flood period build up of the basal accumulation. This results from collapse of bank overhangs and free - falls from the vertical faces (Photo 4.16). The final stage starts with the development of border ice along the river banks in early winter.

The erosion sequence is similar for riparian shale cliffs, till and glaciolacustrine deposits. These deposits usually, but not invariably, have toe accumulations of readily erodible material overlying stable deposits. The sediment supply to the toe deposit is controlled by weathering. Removal is progressive, producing small cliffs similar to the cliffs developed in the bank deposits (Figure 4.18).

Various scenarios of flood generation, which affect the suspended sediment regime at Bragg gauge, can occur because the specific timing of loads depends on the timing and magnitude of tributary runoff in relation to flood generation from the upper basin. For example, if large tributary inputs occur from the foothills, as the result of rainfall in the foothills during a period of limited mountain snow melt, then a load peak will result from flushing of the bank deposits in the contributing streams and in the river reach within the flood runoff generating zone. This may be followed by a second load peak when the upper basin produces the main flood wave. If the spring

Photo 4.16 Toe accumulations are built up following the annual flood

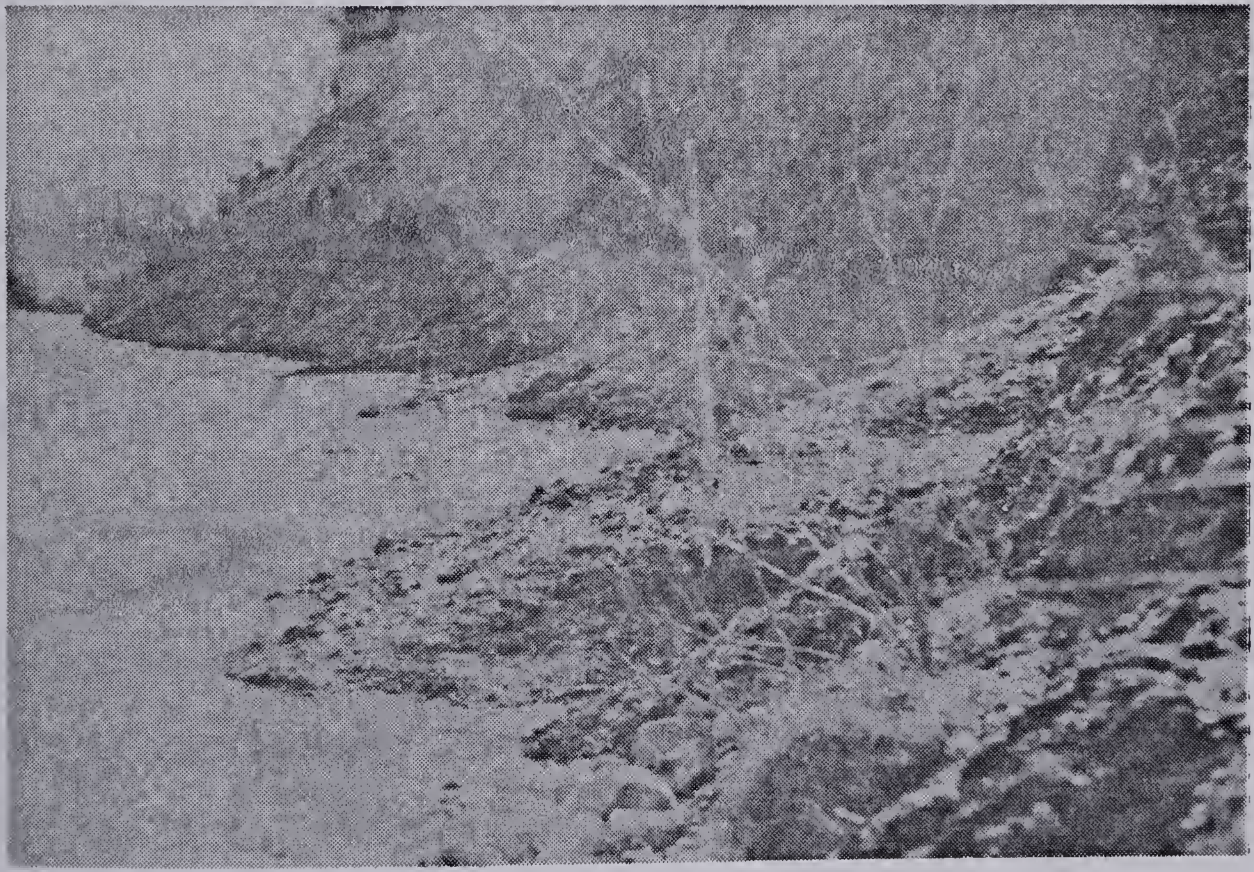


Photo 4.17 Slumping of valley wall deposits in the lower basin is common

flood is generated as the result of significant, general, basin runoff, then one major peak in daily discharge and sediment load results.

In the lower reaches of the Elbow River there are numerous banks which are largely composed of sand and silt. These banks undergo a similar erosion sequence to the largely gravel banks of the upper basin. However, these fine textured banks do not have a toe protected by lag deposits, and, as a result, they erode throughout the high water period, producing a more sustained contribution to the annual suspended load. Protection may be afforded these banks when vegetated overhangs collapse and temporarily remain at the toe (Photo 4.15). In addition, the glacio-lacustrine valley walls may slump directly into the river channel, usually following snowmelt or rainfall (Dawe, 1980). The toe deposits, which constrict the river, are easily removed immediately following slumping, because the deposits are saturated (Photo 4.17).

<u>Comparative magnitude and frequency aspects</u>

Large spatial variations in suspended sediment yield have been recognised at regional and continental scales (e.g. Holeman, 1968). On a world scale suspended sediment concentrations range from close to zero to values in excess of 900 000 mg/l (Gregory and Walling, 1973) and yields in excess of 7000 $t/km^2/yr$ have been reported (Holeman, 1968).

The estimated total sediment delivery to the ocean from North America is about 86 t/km²/yr (Holeman, 1968). This average yield exceeds those of other continents, apart from Asia, which has about six times the average continental yield (Asia 540, Africa 24, Australia 40, Europe 32 and South America 56 t/km²/yr). Within each continent, there are considerable variations in suspended sediment yield. The average drainage basin yield of 20 large North American rivers ranged from 2.8 to 2050 t/km²/yr (Holeman, 1968).

Strakov (1967) (in Gregory and Walling, 1973) estimated the average annual suspended sediment yield of the Eastern Slopes of the Rockies to be about 50 to 100 t/km²/yr. Stichling (1973) estimated the load of the Eastern Slopes foothills and mountains, outside of the National Parks, to be about 18 to 88 t/km²/yr. McPherson (1975) estimated the average suspended sediment yield at Bragg gauge to be 26 t/km²/yr, based on two years of data. Hollingshead et al., (1973) estimated the yield of the Elbow River basin above Glenmore Dam to be 84 t/km²/yr, based on one year of suspended sediment data, a reservoir deposition survey, and an estimate of reservoir trap efficiency. The reservoir trap efficiency has been reviewed in a previous section. The revised yield is 62.5 t/km²/yr.

The estimates of the suspended sediment yield made in this study are in reasonable agreement with the other studies. The major difference in the approach taken here is that spatial variations were examined and yield is shown to

increase substantially downstream from the mountains to the plains, and that there is considerable variation in the yields from individual tributaries and sub - basins. The mountains have an average yield of about 11 t/km^2/yr, the upper foothills 16.3 t/km^2/yr, the lower foothills 34 t/km^2/yr and the plains 138 t/km^2/yr for a basin average of 62.5 t/km^2/yr (Figure 4.19). Luk (1975) estimated sediment yield of the surrounding Bow River basin based on soil erodibility, rainfall, and basin character. He predicted larger yields from the mountains and foothills and lower yields from the plains. The discrepancies may be explained in terms of physiographic differences, sediment delivery, and channel contributions to sediment yield.

The downstream increase in sediment yield in the Elbow River basin contrasts the observation that sediment yield tends to decrease with increasing drainage basin size (e.g. Brune, 1948; Hadley and Schumm, 1961; and Roehl, 1962). Carson et al., (1973) suggest that the increase in sediment yield downstream in the Eaton River basin, Quebec, is explicable in terms of different sources of sediment. Yield tends to increase downstream in drainage basins in which sediment originates primarily from the channel itself. In basins with a large upland surface erosion component, sediment supply (and competence) would be expected to decrease as the basin slope decreases, and usually as basin size increases. Further, with increasing basin size, there is usually an increase in bottomlands, terraces and

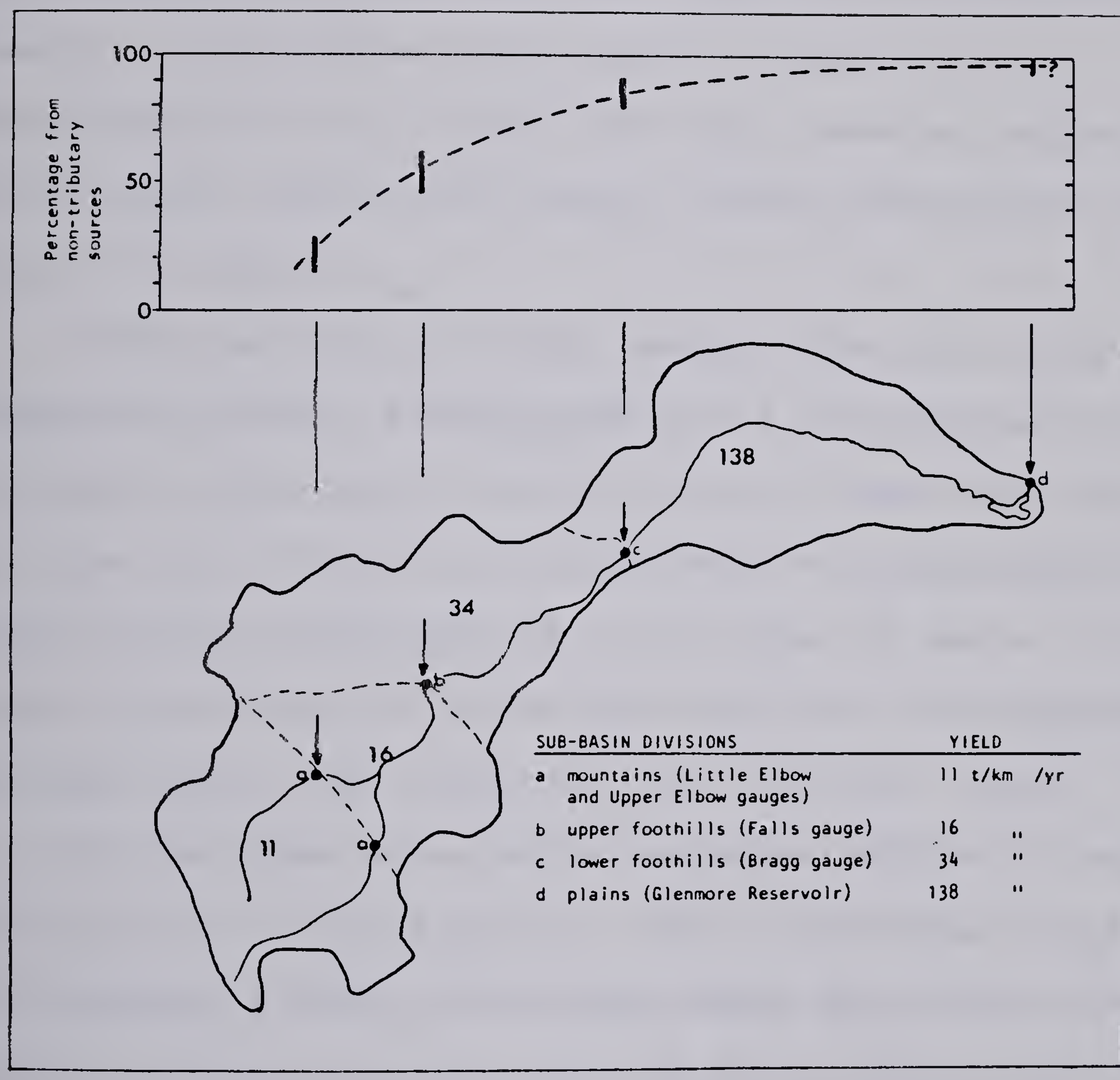


Figure 4.19 Elbow River basin sediment yields and sediment sources - a graphical summary

floodplains, which provide deposition sites for alluvium and colluvium (Hadley and Schumm, 1961).

In the Elbow River basin the downstream increase in sediment yield is the result of increased availability of finer textured material in the lower basin. The low yield of the mountains of the Elbow River basin is attributed to the paucity of available finer material. This may be an explaination of low yields from high elevation basins on a world scale, which is an anamoly Jansen and Painter (1976) could not explain.

Wolman and Miller (1960) suggest that 80 to 95% of the suspended sediment transported over a long period is moved in events which have a return period of less than one year and that 98 to 99% of the total load is transported by events with a return period of less than 10 years. In other words, rare events of great magnitude are not responsible for the bulk of the transport over the long - term.

In the Elbow River basin suspended sediment transport occurs over a limited period - 98% of the annual load occurs in 3 months. Further, on average about 20% of the total annual load occurs in one day and 50% to 80% of the total annual load is transported within 10 days (Figure 4.17).

At Bragg gauge the annual suspended sediment load is expected to exceed about 1 600 t once a year, 38 750 t once every 10 years and about 100 000 t every 100 years (Figure 4.11). The frequency analysis suggests that the majority of the long - term suspended sediment load would be expected to

occur as the result of events with a return period between 1 and 10 years. Smaller, more frequent events transport little sediment, while the largest events occur too infrequently to be of long - term significance. However, this analysis presupposes climatic stability, whereas it is known that the past several decades have been relatively mild (Appendix 2).

E. CONCLUSION

Reservoir sedimentation surveys show that Glenmore Reservoir, near the mouth of the Elbow River, traps about 75600 t of suspended sediment per year from the steep, gravel - bed river. This translates into a unit load of about 62.5 $t/km^2/yr$ from the 1210 km^2 basin. The reservoir deposits are derived almost exclusively from fluvial suspended sediment transport.

Measurement of the suspended sediment load for eight years at Bragg gauge, the mid - point of the system, and measurement of upstream and downstream loads, at 18 points in the system for two years, show that load increases greatly downstream and that the rates of erosion are highly variable in space and time. In the mountains the major sediment sources are upland slopes of barren colluvium. Sediment yield is supply limited in the calcareous bedrock areas because few fines are produced as the result of mechanical weathering. Additional loads, derived from mass movements, particularly of finer textured glacial deposits, into the river and tributary channels, produce local yields more than

twice that of the mountain long - term average yield of 11 $t/km^2/yr$. Sediment trapping causes lower yields from other areas in the mountains.

The load of the upper foothills area is derived almost equally from channel erosion of the tributary streams and main river channel. The dense vegetation cover prevents significant upland erosion. In the lower foothills about 80 to 90% of the estimated long - term load is derived from river channel erosion. Tributary inputs are small, although the area has a high erosion potential, because the dense vegetation cover prevents upland erosion and because numerous large ponded areas, often behind beaver dams, effectively trap most of the suspended sediment load. Accelerated erosion, from roads and trails, occurs in limited areas. The long - term load at the upper foothills hydrometric station, Falls gauge (station 5BJ6), is 6050 t/yr. The load at Bragg gauge in the mid foothills is 18200 t/yr. The long - term load at Sarcee gauge, immediately above Glenmore Reservoir, is about 70,000 t/yr. The long - term load rate of deposition in Glenmore Reservoir averages 75600 t/yr. Reservoir trap efficiency is almost 100%.

In 1978 and 1979 the lower basin load was derived almost exclusively from channel erosion and riparian erosion. It is probable that in years when significiant runoff is generated from the plains area substantial upland erosion of the glaciolacustrine deposits, which mantle the area, will occur. Slumping of valley wall deposits may occur

in the lower basin resulting in high suspended sediment concentrations. Slumping may occur as the result of rainfall or snow melt, and need not be associated with high river discharge.

5. BEDLOAD TRANSPORT

A. INTRODUCTION

Bedload consists of material that forms more than 90% of an alluvial stream bed (Simons and and Sentruk, 1977), which moves by rolling, sliding or saltating on or near the bed (ASCE, 1975). The distinction between bedload and suspended bed material load is dynamic (Church and Gilbert, 1975). Sediment samplers provide an arbitrary, but convenient, definition of bedload and suspended bed material load - bedload samplers measure bedload and suspended sediment samplers measure suspended sediment load (Einstein, 1947).

The objectives of this chapter are to: (a) describe bedload transport at several points within the Elbow River basin (Figure 5.1); (b) compare the findings with selected rivers where bedload has been measured (Table 5.1) and attempt to explain the character of the variations in the bedload; (c) evaluate methods of predicting bedload transport; and (d) predict the long - term rates of transport from each physiographic zone so as to evaluate the relative efficiency of each mode of sediment transport (bed-load, suspended load and solute load) in the fluvial regime.

Table 5.1 Characteristics of several gravel - bed streams in which bed load has been measured

Stream	drainage area km^2	discharge m^3/s	water surface width (m)	average depth or hyd. radius	water surface slope x 10^{-3}	pavement data D90	pavement data D50	discharge (m^3/s) vs load (tonnes/day)	r^2
Clearwater R.	25000	289 - 3512	125 - 149	3.29 - 6.64	0.09 - 0.62	143'	72'	$= 1.175 \times 10^{-6} Q^{2.598}$	0.73
East Fork River	466	2.04 - 45.0	14.8 - 18.0	0.24 - 2.01	$\bar{x}$ - 0.70	4.0	1.0	$= 0.513 Q^{1.566}$	0.51
Elbow R.	795	30 - 109	29 - 34	0.70 - 1.14	7.26 - 7.50	132	76	$= 1.269 \times 10^{-5} Q^{4.156}$	0.41
North Saskatchewan R.	4600	175 - 348	104 - 109	1.4 - 2.5	$\bar{x}$ - 1.58	60	19	$= 7.418 \times 10^{-5} Q^{0.936}$	0.05
'Oak Creek	7	1.16 - 3.40	5.0 - 6.1	0.31 - 0.45	9.70 - 10.8	86	54		
Snake R.	240000	780 - 4670	155 - 204	3.26 - 6.77	0.56 - 1.40	137'	54'	$= 1.175 \times 10^{-5} Q^{2.23}$	0.40
Vedder R.	1227	95 - 372	70 - 100	1.4 - 3.3	$\bar{x}$ - 1.95	90'	44'	$= 2.427 \times 10^{-6} Q^{3.023}$	0.35

('From Parker et al., 1982)
hydraulic characteristics over the range of bedload measurements, except for Oak Creek

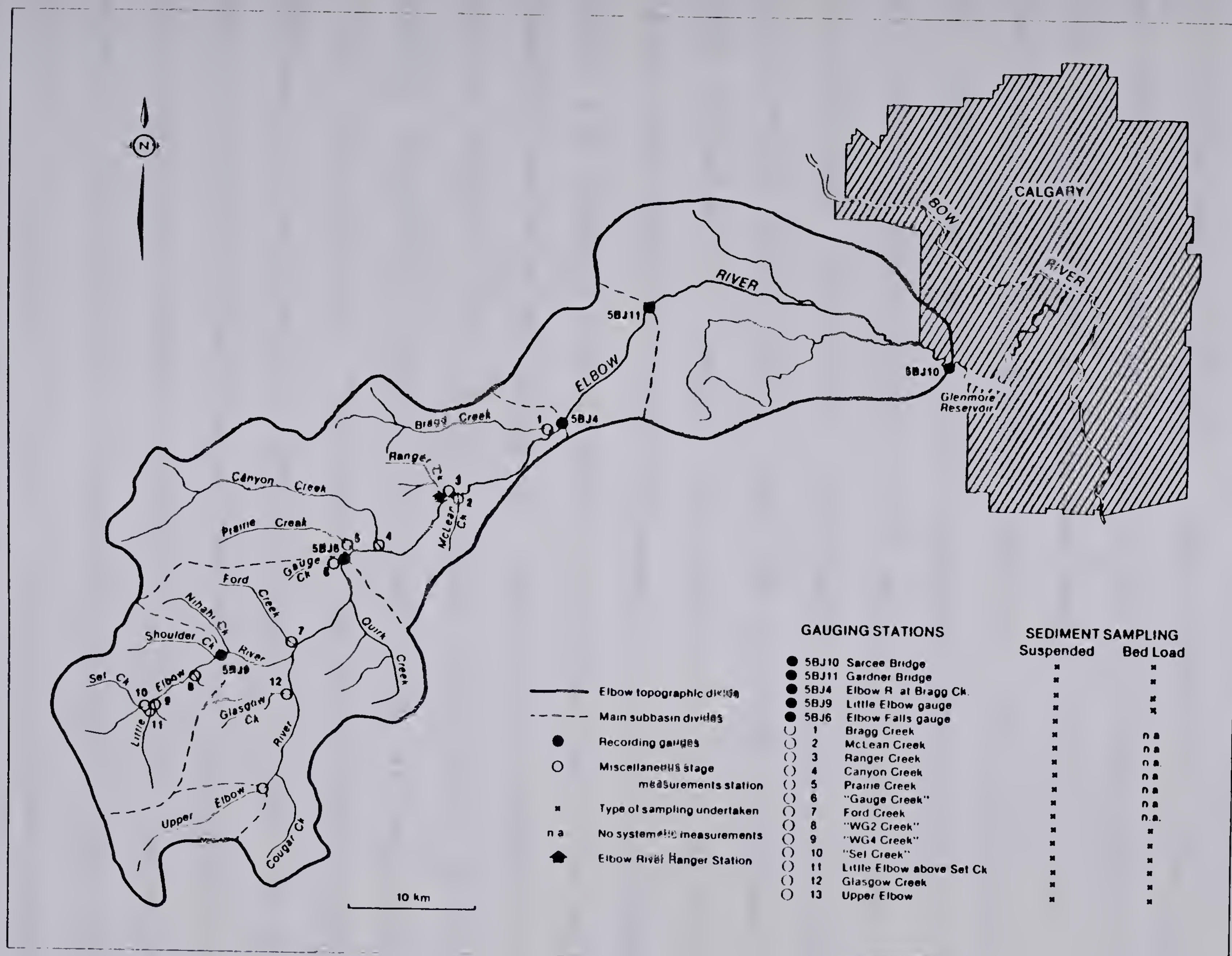


Figure 5.1 Elbow River basin hydrometric and sediment sampling stations

B. METHODOLOGY

A Helley - Smith 75 mm sampler with a large one mm mesh sample bag, was made by the writer for this project. The device is a modified version of the Arnhem pressure difference sampler (Emmett, 1980) (Photo 5.1). The sampler was held on the bed by a wading rod or a long heavy pipe (8 m long, 4 cm diameter), which was usually held at the downstream side of a bridge. The lower section of pipe above the sampler was guyed to the upstream side of the bridge. With practice the sampler could be seated on the bed, even in high flows, without any perceptible gouging or slipping motion which would disturb the bed or cause an excessive catch of material. Samples were taken at 10 to 20 verticals across the cross sections (Figures 5.2 to 5.5). The sampling period was usually two minutes at a point but this varied, depending on rate of transport, from thirty seconds to five minutes. Emmett (1976) used a sample duration of 30 to 60 seconds. Replicate sampling was undertaken at each station in the cross section. The sampler was positioned slightly upstream of the previous sampling point, in case the earlier sampling had disturbed the stream bed. If the amount caught was inconsistent between samples at a station, the whole procedure was repeated. If measured transport rates were consistent, but quite different across the stream, samples were taken at intermediate locations, in a attempt to delineate the zone of discontinuity.

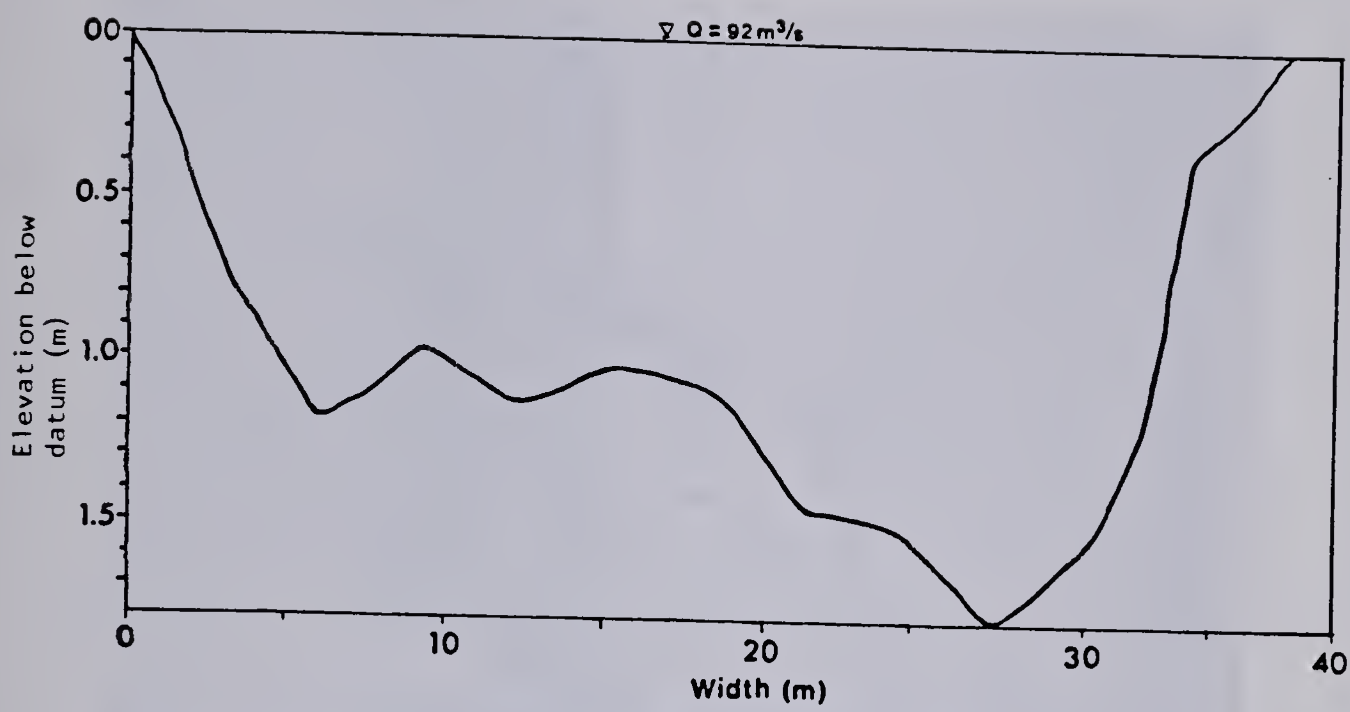


Figure 5.2 Elbow River at Bragg gauge, cross - section

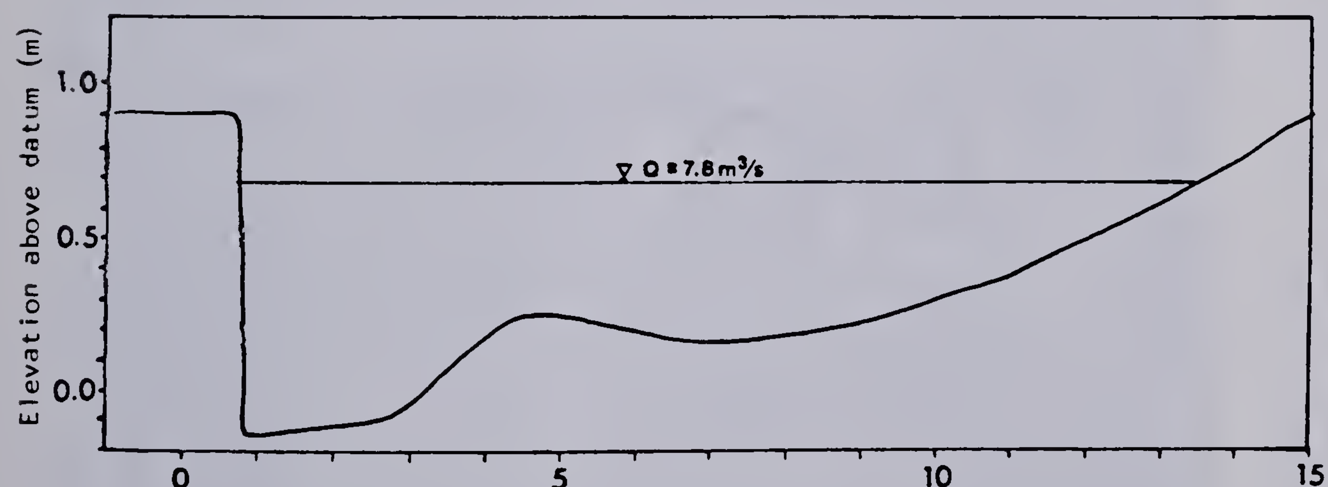


Figure 5.3 Little Elbow River gauge, cross - section

Photo 5.1 The wading rod version of the 75mm Helley-Smith bedload sampler

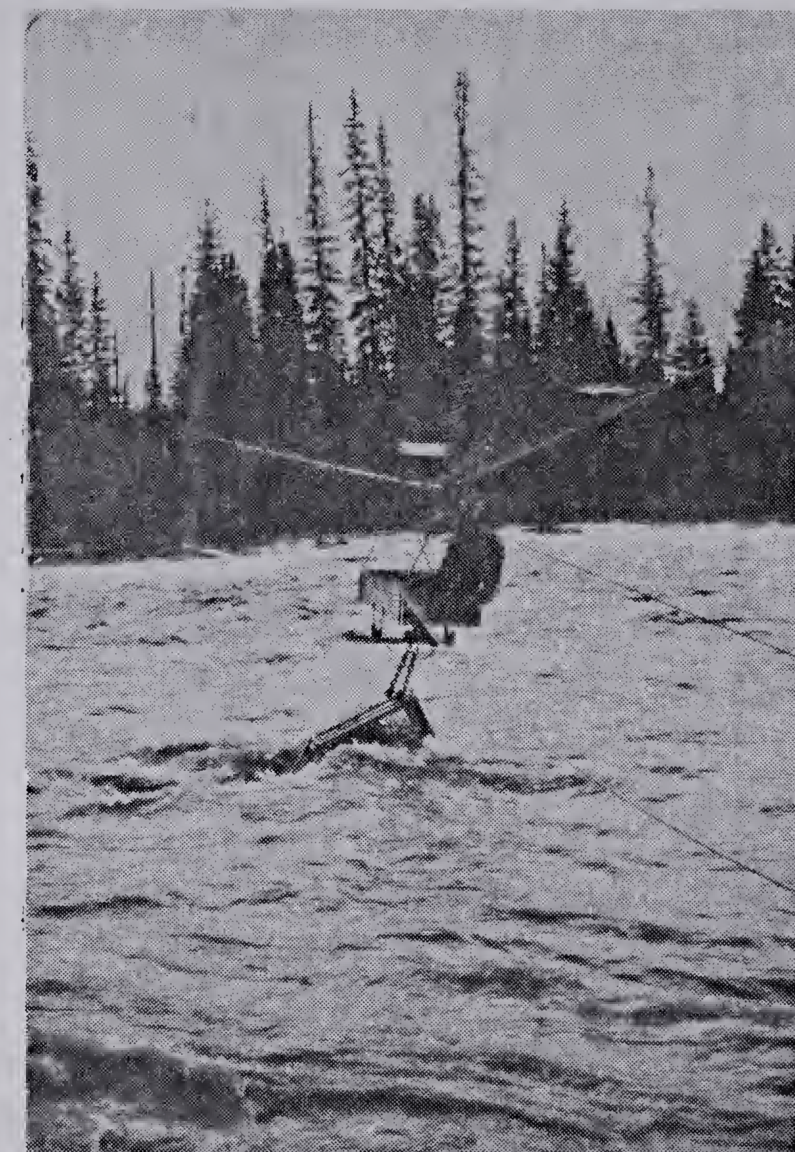


Photo 5.2 A large basket sampler was used for bedload measurements near Bragg gauge in the 1967 to 1979 period

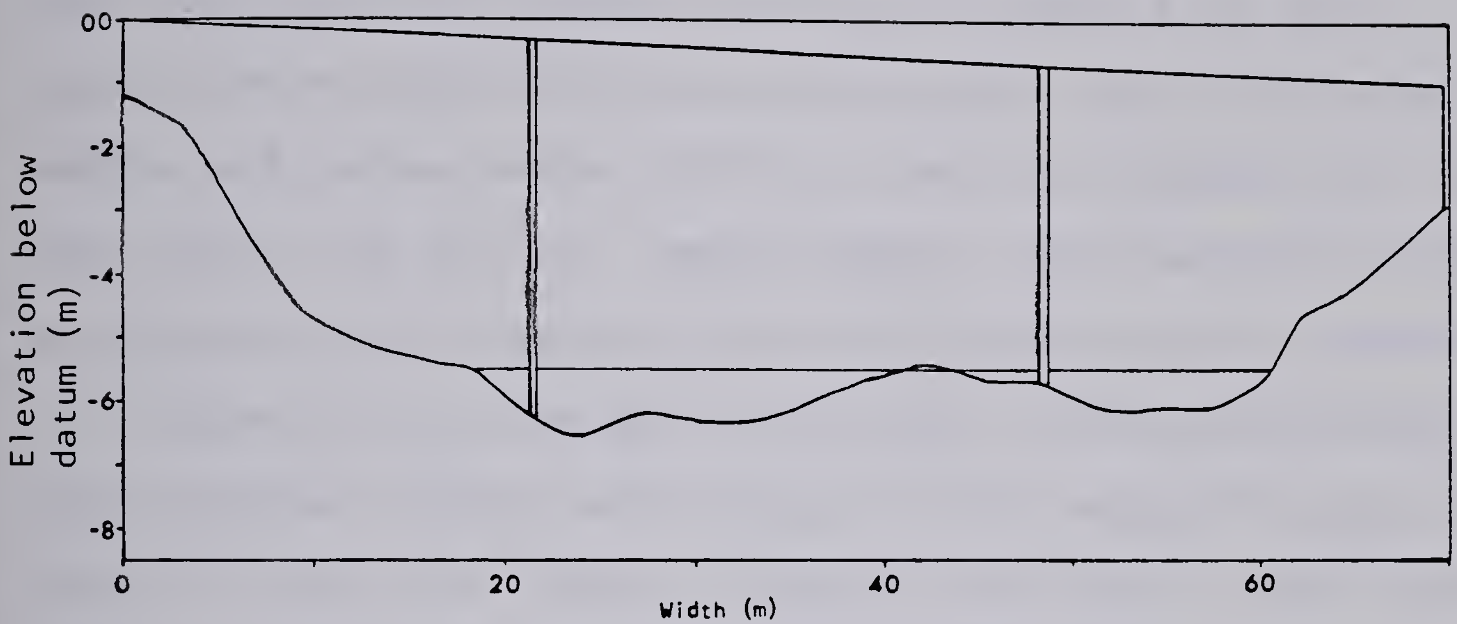


Figure 5.4 Elbow River at Gardener gauge, cross - section

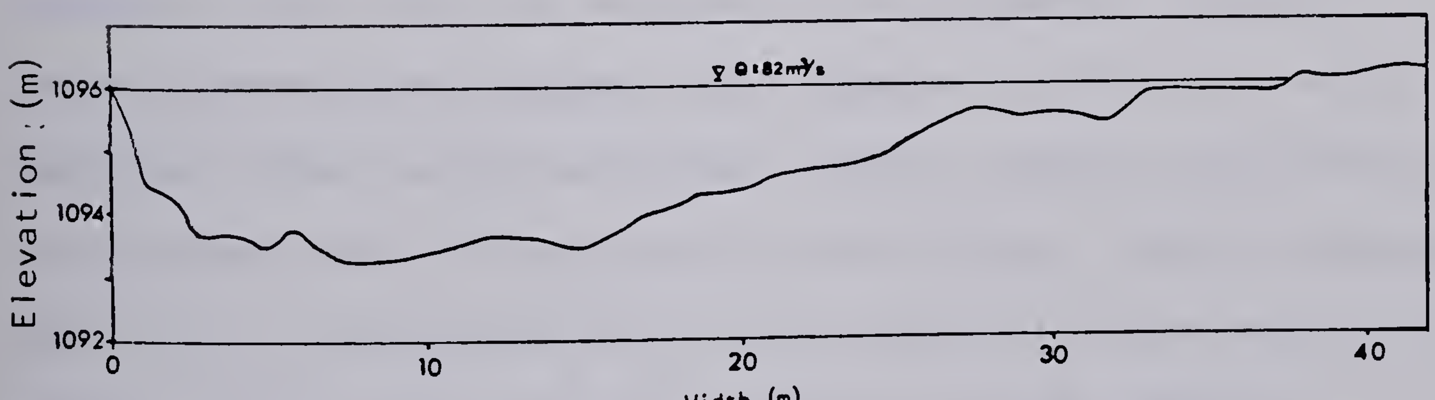


Figure 5.5 Elbow River at Sarcee gauge, cross - section

Bedload samples were individually bagged, air dried and sieved at half phi intervals. Material less than 0 phi (1 mm) was usually not sieved to further differentiate grain size distribution. However, several of these fine sub - samples were sieved and the results demonstrate that some medium and coarse sands (0.25 to 1 mm) are trapped by a 1 mm mesh bag on the Helley - Smith sampler. This material formed an insignificant portion of the total weight of the sample.

There are limited data with which to evaluate Helley - Smith bedload sampler efficiency for the range of sizes in motion in the Elbow River. Klingeman and Emmett (1980) state that the 3 inch (75 mm), cable - supported, Helley - Smith sampler used at the East Fork River, had a 100% sediment trapping efficiency, for material from 0.5 to 16 mm. For larger particles the calibration data were insufficient to establish a sampling efficiency. Observations in clear, shallow mountain streams in the Elbow River basin suggest that particles which are almost the size of the sampler entrance move into the sampler in an unimpeded fashion. These observations suggest that sampler efficiency would be high for these large particles. Hence, sampler efficiency corrections were not applied to the Helley - Smith bedload data. It is recognised that up to 20% of the bed material in motion is too large to be caught in a 75 mm sampler. However, a correction for this exclusion would not shift the data points significantly (Hudson, in press (a)).

Hollingshead (1968) and Samide (1971) used various samplers to measure bedload on the Elbow River at Bragg Creek. The most frequently used sampler was a large basket, with a 6.4 mm (1/4") mesh (Photo 5.2). In addition VUV, 12.5 and 19 mm mesh basket samplers were used in 1967, 1968 and 1969. Hollingshead (1968) assumed an efficiency of 70% for the VUV sampler, and 45% for the 6.4 mm mesh basket sampler. In a subsequent paper (Hollingshead, 1971) he modified his basket sampler efficiency, by suggesting that the sampler was on average 45% efficient (based on Hubbell, 1964), but given that trapped grain sizes were coarser than pit samples, he added a further factor to arrive at a much larger correction: 1/(0.45 X 0.70) = 3.18. He then multiplied all his data by 3.18 to arrive at an estimate of sediment discharge. Subsequent work by Samide (1971), who calibrated the samplers in the Elbow River, suggests that the grain size trapped by the 6.4 mm mesh sampler is not significantly different from those caught by the VUV sampler, which was used concurrently. Hubbell (1964:71) states: "The principal advantage of the VUV sampler is that it usually gives a good representation of the actual bedload size distribution". This appears to be the case in the Elbow River (Figure 5.6). Further, Samide (1971) proposed a progressive increase in sampler efficiency with discharge up to about 45 m^3/s, when an efficiency of 45% should be used. Gibbs (1973) concurs with this finding.

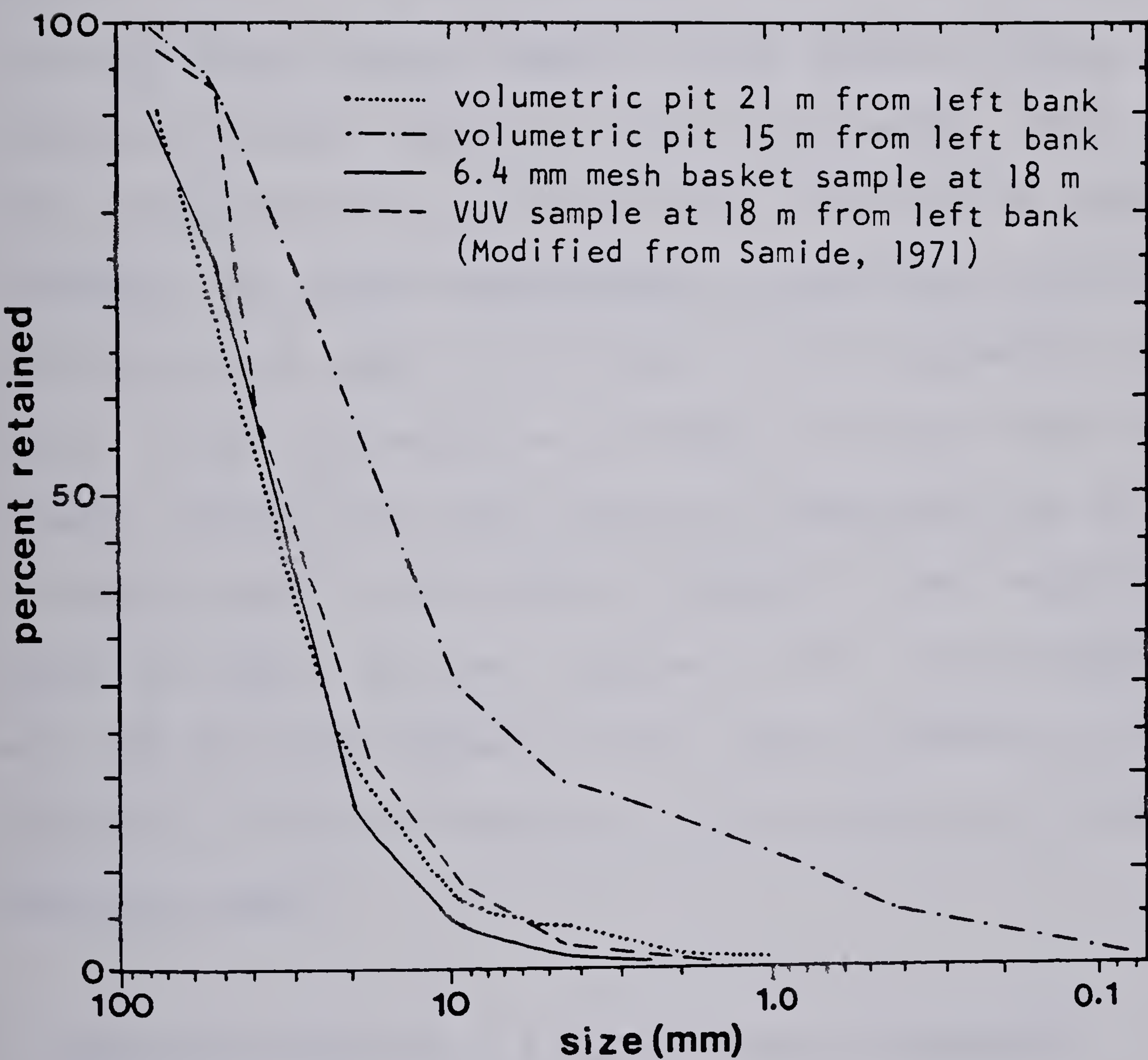


Figure 5.6 Bed material and bedload sampler grain size distributions, Elbow River at Bragg gauge

The above analysis suggests that Hollingshead (1971: his Table 1) has over - corrected for sampler efficiency, and that he should have used a factor of about 2.22, rather than 3.18, to estimate bedload from the sampler data, for the range of conditions sampled. Further, the original data bases were examined (Hollingshead, 1968 and Samide, 1971) and it was found that three types of sampler were used to produce Table 1 of Hollingshead (1971), not just a 1/4" (6.4 mm) large basket sampler as he states. Hence, different efficiency values should be applied (Samide, 1971; Gibbs, 1973) to arrive at a more accurate estimate of bedload transport for these measurements. Additional corrections and additions were made to the data in Hollingshead's (1971) Table 1. The revised data in Table 5.2 show that Hollingshead's (1968) data was correctly reported, but the estimated load is too great, because of the correction factor he used. However, Samide's (1971) data summary, which was used by Hollingshead (1971), has a number of errors and omissions, notably regarding active widths of transport and measured loads.

C. BEDLOAD TRANSPORT - A DESCRIPTION OF RESULTS

Introduction

A significant feature of the Elbow River bedload transport is its variability at a number of scales, in both time and space. This topic is the subject of the following

Table 5.2 Elbow River at Bragg gauge, 1967, 1968 and 1969

hydraulics and bedload, revised data summary

Date	Q m^3/s	$\bar{v}$ m/s	d m	w m	s m/m	ω N/m^2	T N/m^2	Load kg/hr	Load tonnes/day	aw m	d_{50} mm	max. size mm	timing wrt. Q max	
1-6-67	109	2.66	1.13	34.4	0.00745	220	83	44988	1080	18.3		185	A	—
2-6-67	82	2.32	1.02	33.1		173	75	23542	565	15.2		96.5	A	F
5-6-67	52	1.87	0.86	31.1		118	63	24.5	0.59	6.1		25.4	A	F
17-6-67	54	1.90	0.88	31.2		122	64	27543	661	24.4		127	A	F
19-6-67	54	1.90	0.88	31.2		122	64	37068	890	21.3		88.9	A	F
19-6-67	47	1.78	0.83	30.6		108	61	2531	61	9.1		102	A	F
20-6-67	46	1.76	0.82	30.5		105	60	5171	124	12.2		152	A	F
21-6-67	44	1.72	0.81	30.3		102	59	12655	304	18.3		109	A	F
22-6-67	42	1.68	0.80	30.1		98	58	4708	113	15.2		76	A	F
22-6-67	42	1.68	0.80	30.1		98	58	11376	273	15.2		127	A	F
23-6-67	35	1.54	0.75	29.4		84	55	161	3.9	9.1		50.8	A	F
9-6-68	40	1.64	0.78	29.9		93	57	4518	108	13.7			—	—
25-6-69	83	2.34	1.03	33.2		176	25	91119	2187	12.2			B	F
26-6-69	91	2.44	1.06	33.6		189	77	53697	1289	12.2	31.		B	F
27-6-69	66	2.09	0.94	32.1		144	69	69673	1672	21.3	51.		B	F
28-6-69	63	2.05	0.93	31.9		139	68	43736	1050	17.5	53.		B	steady
30-6-69	91	2.44	1.06	33.6		189	77	79825	1916	15.2	51.		A	F
1-7-69	84	2.35	1.03	33.2		177	75	60283	1447	12.2	76.		A	F
2-7-69	69	2.14	0.96	32.3		150	70	60229	1445	16.0	40.		A	F
3-7-69	61	2.01	0.92	31.8		135	67	16738	402	15.2	36.		A	F
4-7-69	59	1.98	0.90	31.6		130	66	13662	328	18.3	32.		A	steady
5-7-69	57	1.95	0.89	31.5		127	65	7920	190	18.3	56.		A	F

where : Q = mean discharge (m^3/s) over the bedload sampling period;
V = mean velocity (m/s); d = mean depth (m); w = surface water width (m);
S = average water surface slope for the reach; ω = unit stream power (the product of the unit weight of water, v,d and s; T = shear stress at the bed (T = ω/v); a.w. = active width of bedload movement (m); D_{50} and max. size are the median and largest particle sizes trapped, respectively; timing: A is a measurement following, and B is a measurement before the annual maximum discharge; R and F are rising and falling stage measurements, respectively.
1967 and 1968 data from Hollingshead (1968)
1969 data from Samide (1971)

sections.

Temporal variations in bedload

The relationships between bedload transport and discharge for the Elbow hydrometric stations are illustrated in Figure 5.7 to 5.10. Although there is considerable scatter in the relationship between bedload and discharge, bedload tends to be proportional to about the fourth or fifth power of the discharge. The exponents for several other rivers range from 0.9 to 4.1 (Table 5.1). The large exponent illustrates the non - linear nature of bedload transport and suggests that the majority of transport would normally occur during a limited period of high discharges. Hollingshead (1971: 1833) indicates that at Bragg gauge "The bed is at state of incipient movement at a flow of about 800 cfs (22.7 m^3/s), which is exceeded about 30 days per year on average".

During the period in which threshold conditions are apparently exceeded, very large fluctuations in bedload discharge occur for similar hydraulic conditions. There does not appear to be a systematic variation in loads between the rising and falling limbs of the hydrographs (Figure 5.7 to 5.10), perhaps because the water surface slope remains relatively constant from low stage to high water (7.26 to 7.50×10^{-3} with an average of 7.45×10^{-3} at Bragg gauge; Hollingshead, 1971). There may be a considerable variation in the measured transport rate on different occasions, for

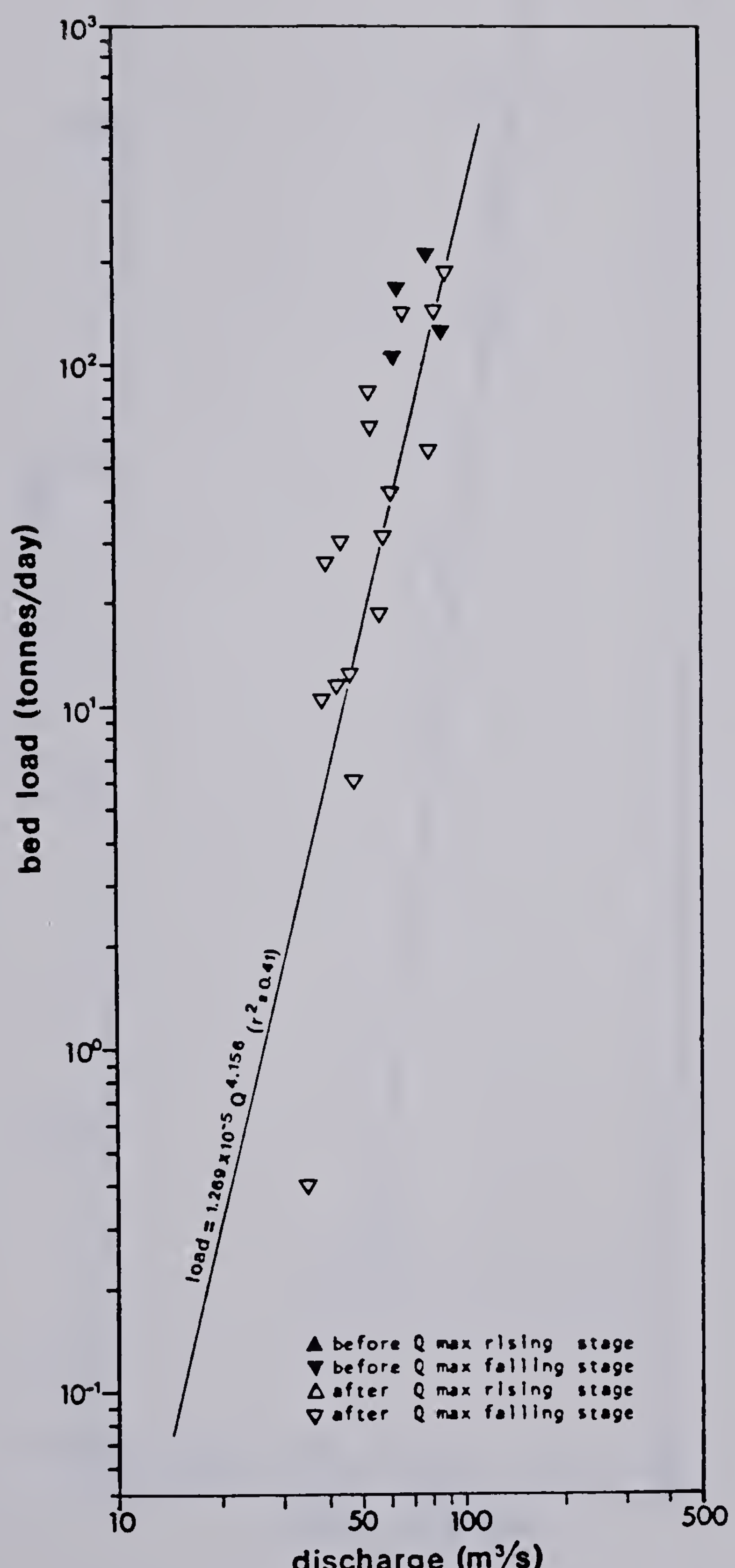


Figure 5.7 Revised bedload data against discharge, Elbow River at Bragg gauge

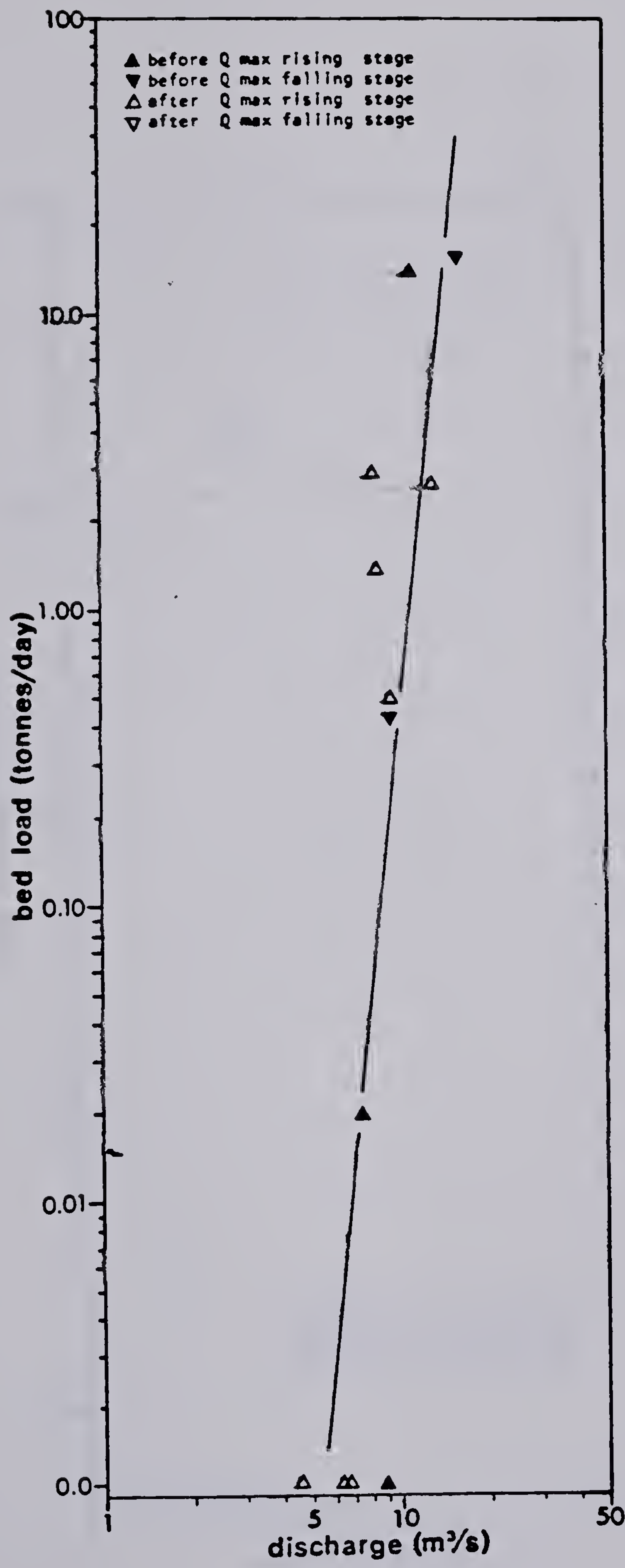


Figure 5.8 Bedload against discharge, Little Elbow River gauge

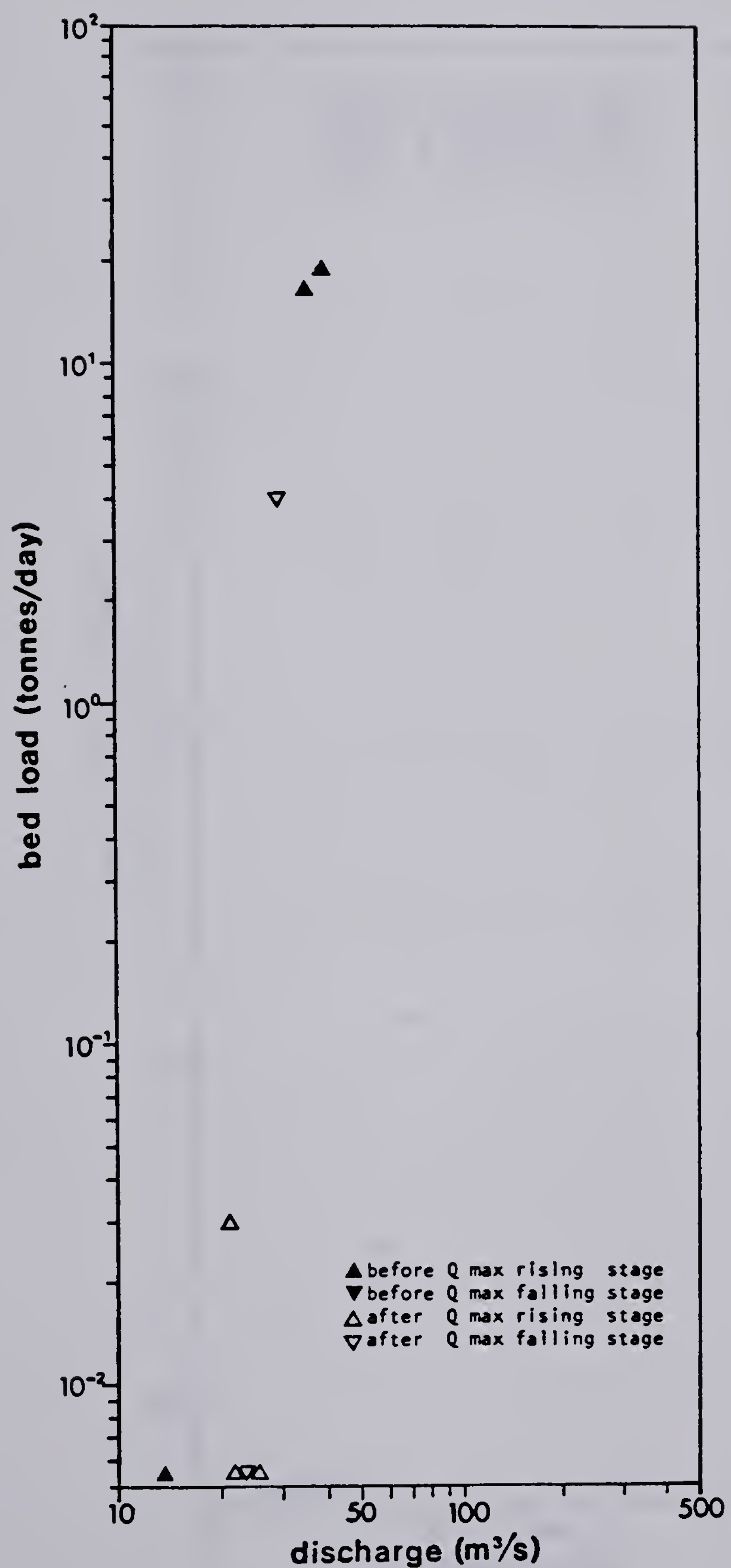


Figure 5.9 Bedload against discharge, Elbow River at Gardener gauge

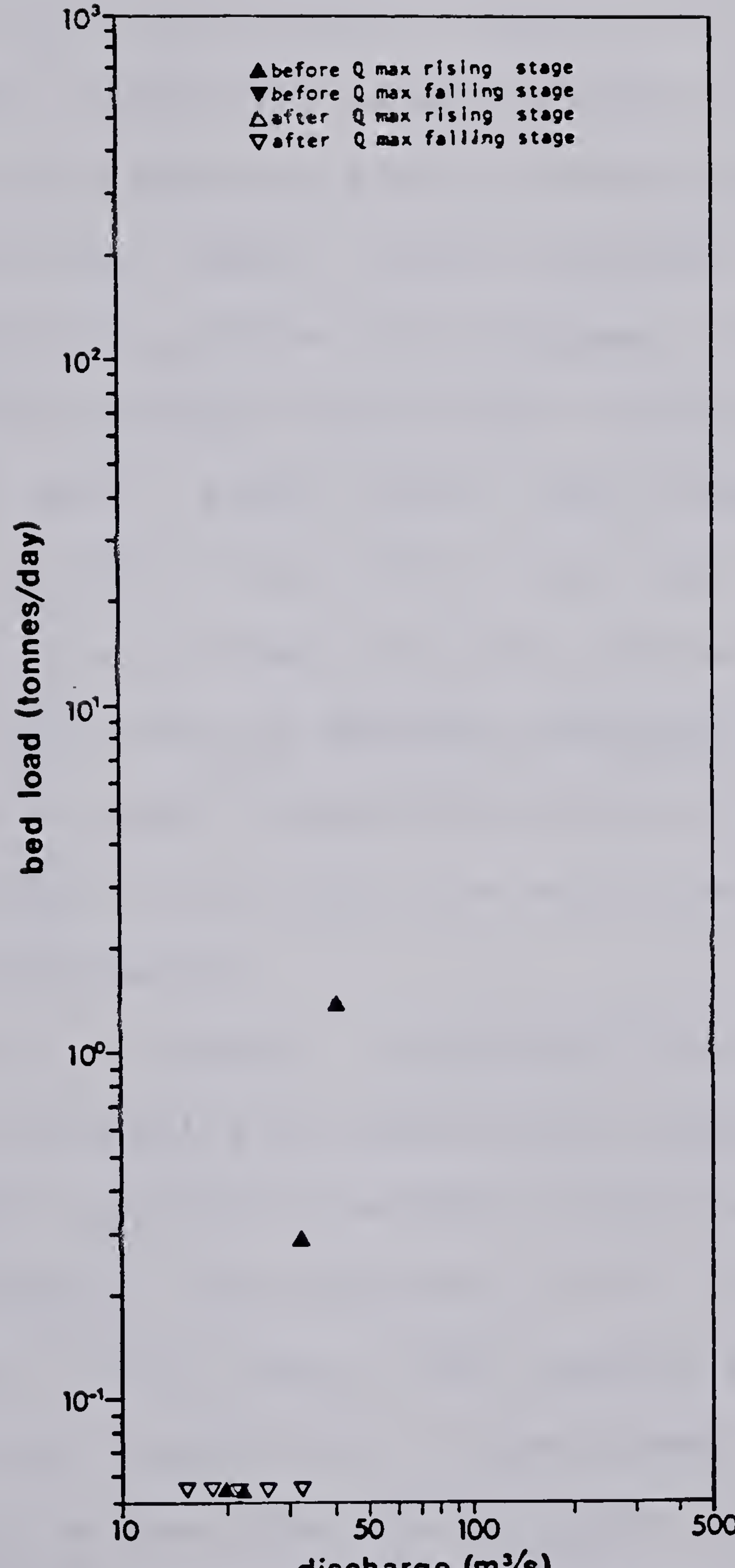


Figure 5.10 Bedload against discharge, Elbow River at Sarcee gauge

similar conditions, e.g.: Q = 52 m^3/s, load = 0.59 t/day; Q = 54 m^3/s load = 890 t/day (Table 5.2).

The large difference in loads for similar hydraulic conditions can not be explained by errors in sampling, i.e. too short a period of sampling with respect to the inherent periodicity of bedload movement (Figure 5.11) (Hudson, in press (a)). The variability is attributed to (a) sediment supply - limited transport, and (b) variability due to the mechanics of motion. Supply limited transport has two components. Firstly, Milhous and Klingeman (1971) showed that stream bed pavements[3] controlled the rate of bedload transport in a small, steep, gravel bed stream. Second, Nanson (1974) and Griffiths (1979) state that contributions from upland erosion and bank erosion produced large variations in the rates of bedload transport in headwater streams. These streams transported bedload at less than capacity rates before and after the exotic material was flushed from the channel.

In addition to supply - constrained bedload transport, considerable variability in the rate of transport may occur at a point over a period of seconds to minutes in rivers and flumes (Ehrenberger, 1931; Einstein, 1937; Samide, 1971; Gibbs and Neill, 1973; Jonys, 1976; Leopold and Emmett, 1976) (Figures 5.11 and 5.12). In some cases, such as the East Fork River at low flow, the irregular cyclic variations

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[3]The term pavement is used rather than armour. Pavement is a coarse surface layer of particles on the stream bed which move during periods of high flows. Armour is bed material which "never" moves (Parker et al., 1982).

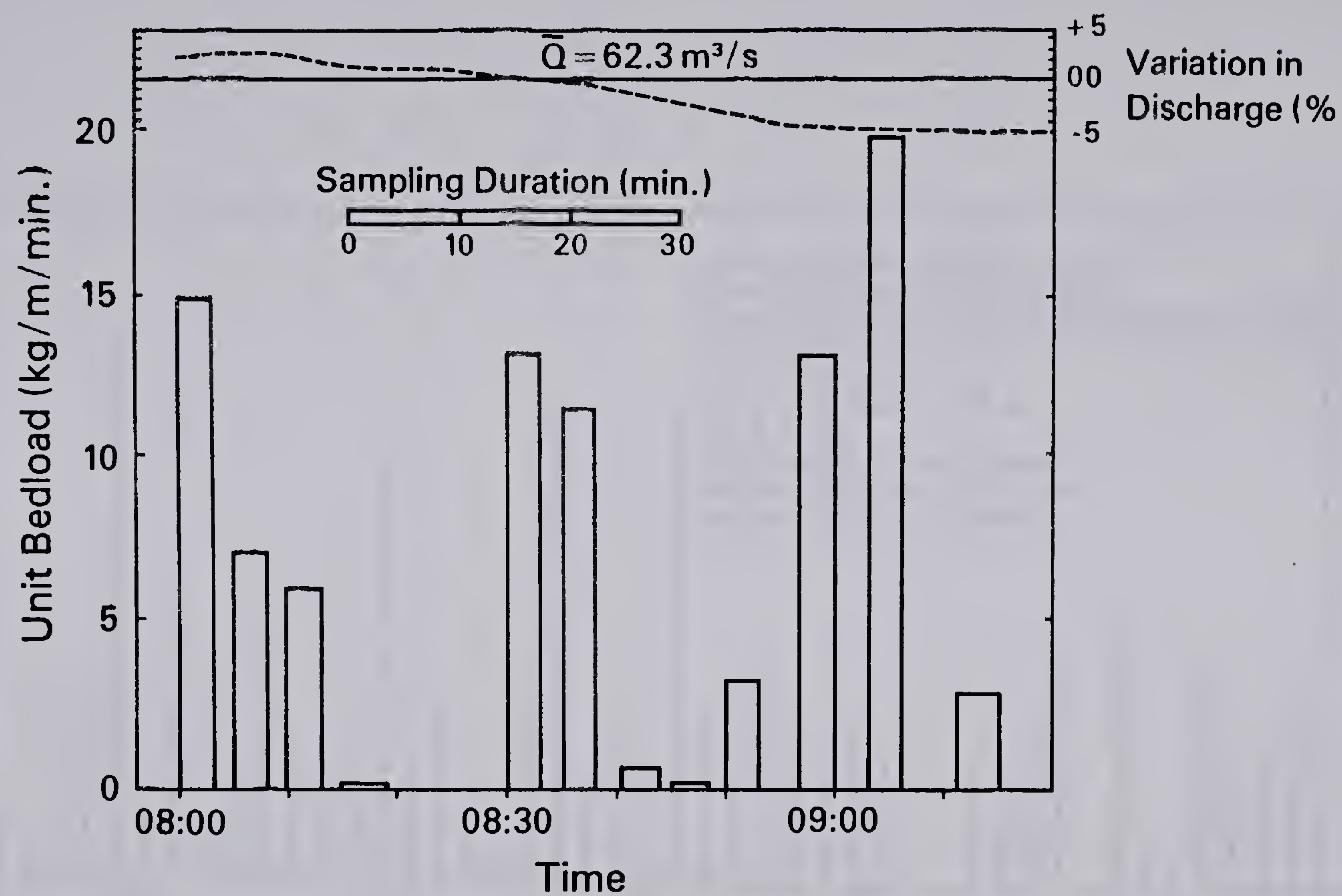

Figure 5.11 Bedload variations with time, at a fixed point, Elbow River at Bragg gauge (modified from Samide, 1971)

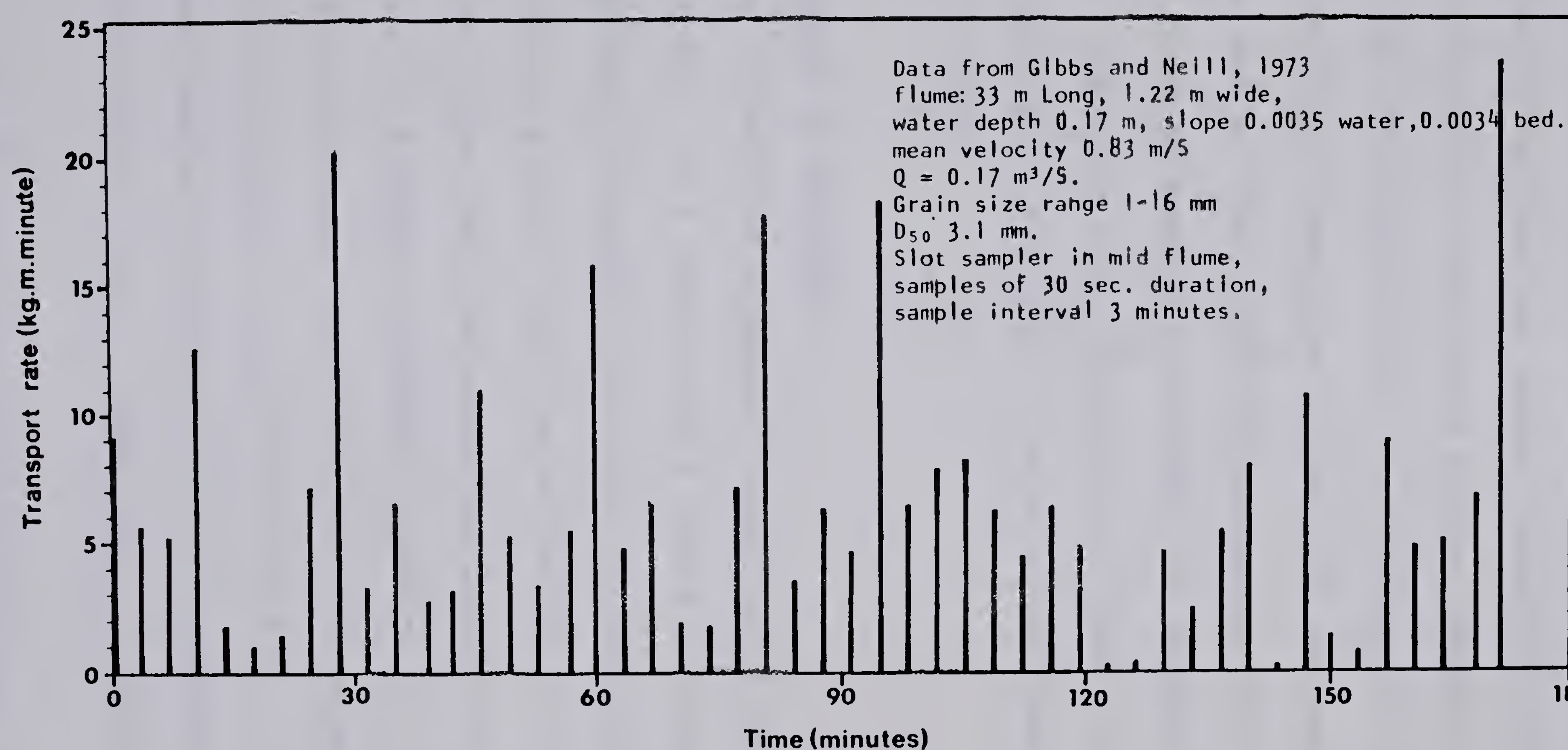

Figure 5.12 Bedload variations with time in a large, gravel - bed, flume

are due to bedload moving as sand dunes over a gravel pavement (Leopold and Emmett, 1976). However, at high stage in the East Fork River, movement occurs in sheets or tongues not associated with any obvious dunes (Leopold and Emmett, 1976). The pulsing is not explained by the passage of dunes in the Elbow River. The variability is thought to be due to the passage of diffuse sheets or carpets of bed material moving over the bed as short period kinematic waves (Hudson, in press (a)). Such movements are common in deserts (Bagnold, 1941) and have been reported in flumes (Langbein and Leopold, 1968) and in rivers (Leopold and Emmett, 1976; Hein and Walker, 1977).

Spatial variations in bedload

Variations in bedload transport occur at a meso - scale and a micro - scale. At a basin scale there is considerable variation in bedload transport from the tributaries and within the river, and there is a major change in the rate of bedload transport from the mountains downstream to Sarcee gauge (Figure 5.1). Bedload transport appears to increase from the mountains through the foothills, and appears to decrease through the plains (Tables 5.2 to 5.4). Load variations downstream are considered in a following section, describing prediction of long - term rates of bedload transport.

Only a few of the mountain tributaries were observed to have gravel bedload transport (Figure 5.1; Table 5.5). Other

Table 5.3 Little Elbow River gauge, 1978 and 1979 hydraulics and bedload data summary

Date	Time ±30 min	Q m^3/s	$\bar{v}$ m/s	d m	w m	s m/m	ω N/m^2	T N/m^2	Load kg/hr	Load tonnes/day	aw m	d_{50} mm	max. size mm	timing wrt. Q max	
22-5-78	1700	8.84	1.57	.47	12.0	1.45×10^{-2}	105	67	—	0.0	—	—	—	B	R
3-6-78	1710	7.62	1.46	.45	11.7		93	64	0.62	0.016	5.0	4.9	9.5	B	R
3-6-78	2330	11.24	1.77	.51	12.6		128	73	591.	14.2	8.1	24.0	67.	B	R
4-6-78	0800	9.67	1.64	.49	12.2		114	70	18.6	.45	1.25	28.0	2.	B	F
4-6-78	2200	16.37	2.12	.58	13.5		175	83	662.	15.9	9.8	25.0	53.	B	F
22-6-78	1630	8.18	1.51	.46	11.9		99	65	59.	1.42	11.4	10.0	26.	A	R
22-6-78	2130	12.80	1.88	.53	12.8		142	75	112.	2.69	9.5	6.8	31.	A	R
25-5-79	2305	9.01	1.59	.48	12.1		109	68	23.5	.56	2.0	3.5	8.	B	R
2-6-79	2125	6.75	1.38	.43	11.5		84	61	T		—	—	—	A	R
3-6-79	2400	8.18	1.51	.46	11.9		99	65	126.	3.02	3.7	22.0	38.	A	F
10-6-79	2200	4.60	1.14	.38	10.7		62	54	—	0	—	—	—	A	steady
12-6-79	2245	6.59	1.36	.43	11.4		83	61	T		—	—	4.	A	R

See Table 5.2 for expla nation

Table 5.4 Gardener gauge and Sarcee gauge, 1978 and 1979 hydraulics and bedload data summary

Date	Time	Q m^3/s	$\bar{v}$ m/s	d m	w m	s m/m	ω N/m^2	τ N/m^2	Load kg/hr	Load tonnes/day	aw m	d_{50} mm	max. size mm	timing wrt. Q max	
Gardener															
24-5-78	1930	13.5	1.00	0.39	32.8	7.07×10^{-3}	27	27	0.0	0.0				B	R
5-6-78	0330	36.2	1.72	0.48	40.8		57	33	7166	172	9.1	34.5	71	B	R
11-6-78	2000	29.0	1.53	0.46	38.8		49	32	1624	39	27.4	28.0	72	A	F
21-6-78	1530	23.2	1.35	0.44	37.0		41	31	0.0	0.0				A	F
27-5-79	0730	40.8	1.85	0.50	41.9		64	35	7832	188	13.9	13.0	84	B	R
3-6-79	0715	21.5	1.30	0.43	36.4		39	30		0.29				A	R
4-6-79	0740	24.7	1.40	0.45	37.5		44	31	12.1	0	3.0	4.1	20	A	R
5-6-79	0700	23.6	1.37	0.44	37.1		42	31	0.0					A	R
Sarcee															
30-5-78	1030	17.0	.84	0.80	25.6	1.776×10^{-3}	12	14		0				B	R
11-6-78	1420	32.2	.97	1.20	28.2		20	21	120	2.83	5.9	11.0	32	A	F
21-6-78	1400	22.2	.89	.95	26.6		15	17		0				A	F
24-6-78	1315	26.0	.92	1.05	27.3		17	18		0				A	F
7-7-78	1210	17.9	.85	0.82	25.8		12	14		0				A	steady
20-5-79	1100	15.3	.82	0.74	25.2		11	13		0				B	R
27-5-79	1010	38.8	1.01	1.35	29.0		24	24	549	13.2	10.0	9.4	18	B	R
3-6-79	0900	18.9	.86	0.85	26.0		13	15		0				A	R
4-6-79	1020	22.4	.89	0.95	26.7		15	17		0				A	R

See Table 5.2 for expla natlon

Table 5.5 Mountain tributaries, 1978 and 1979 hydraulics and bedload data summary

Date Stream	Time	Q m^3/s	$\bar{v}$ m/s	d m	w m	s m/m	ω N/m^2	τ N/m^2	Load kg/hr	Load tonnes/day	aw m	d_{50} mm	max. size mm
WG2													
22-5-78	1800	0.321	.81	.09	2.80	1.3×10^{-2}	9.3	11.5	1.0	0.02	0.50	11.3	11.3
3-6-78	1945	0.712	1.01	.24	3.0		30.9	30.6	496.	11.9	1.90	19.1	19.1
4-6-78	0000	0.696	1.32	.19	2.75		32.0	24.2	178.	4.3	1.65	15.9	15.9
6-6-78	0845	0.356	.83	.16	2.95		17.0	20.4	0		0	0	0
2-6-79	1930	0.447	1.00	.16	3.00		20.4	20.4	0		0	0	0
3-6-79	2115	0.520	1.32	.16	3.00		27.0	15.5	0		0	0	0
WG4													
3-6-78	1945	1.327	1.43	.17	5.70	2.9×10^{-2}	69.2	48.4	370.	8.9	3.05	21.0	53.
6-6-78	0915	0.855	1.03	.16	5.25		46.9	45.5	0		0	0	0
3-6-79	1945	.454	0.74	.13	4.76		27.4	37.0	35.6	0.85	3.0	15.0	20.

major streams in the Upper Elbow River basin, such as Cougar Creek, were inaccessible. The gravels in the foothills streams were stable in 1978 and 1979. No movement of material larger than sand size (2 mm) was observed. The plains tributaries flow through glaciolacustrine sand, silt and clay.

Two of the largest mountain tributaries, Set Creek and the Little Elbow River above Set Creek, which have bouldery beds in their lower reaches, were not observed to transport gravel bedload. For the case of Set Creek, this is attributed to upstream debris jams trapping the bed and suspended loads. In the Little Elbow River above Set Creek, the low loads are attributed to sediment supply - limitations. Deposition behind a large, stable, debris jam suggests that bedload transport has been limited for a period of at least 10 to 15 years.

The two major bedload producers in the Little Elbow River basin, in 1978 and 1979, were WG2 and WG4 (Table 5.5). This situation also appears to be the case for the long - term. The only additional major sediment sources, in the Little Elbow River basin, appear to be two areas of alluvial fan truncation near WG2, and a slump of morainic material, immediately downstream of WG2 (Figure 5.13). Streams WG2 and WG4 originate in steep basins, which are mantled with coarse textured colluvium derived from limestone and dolomite.

At a micro - scale, there is a limited width over which bedload movement occurs within a cross - section. The

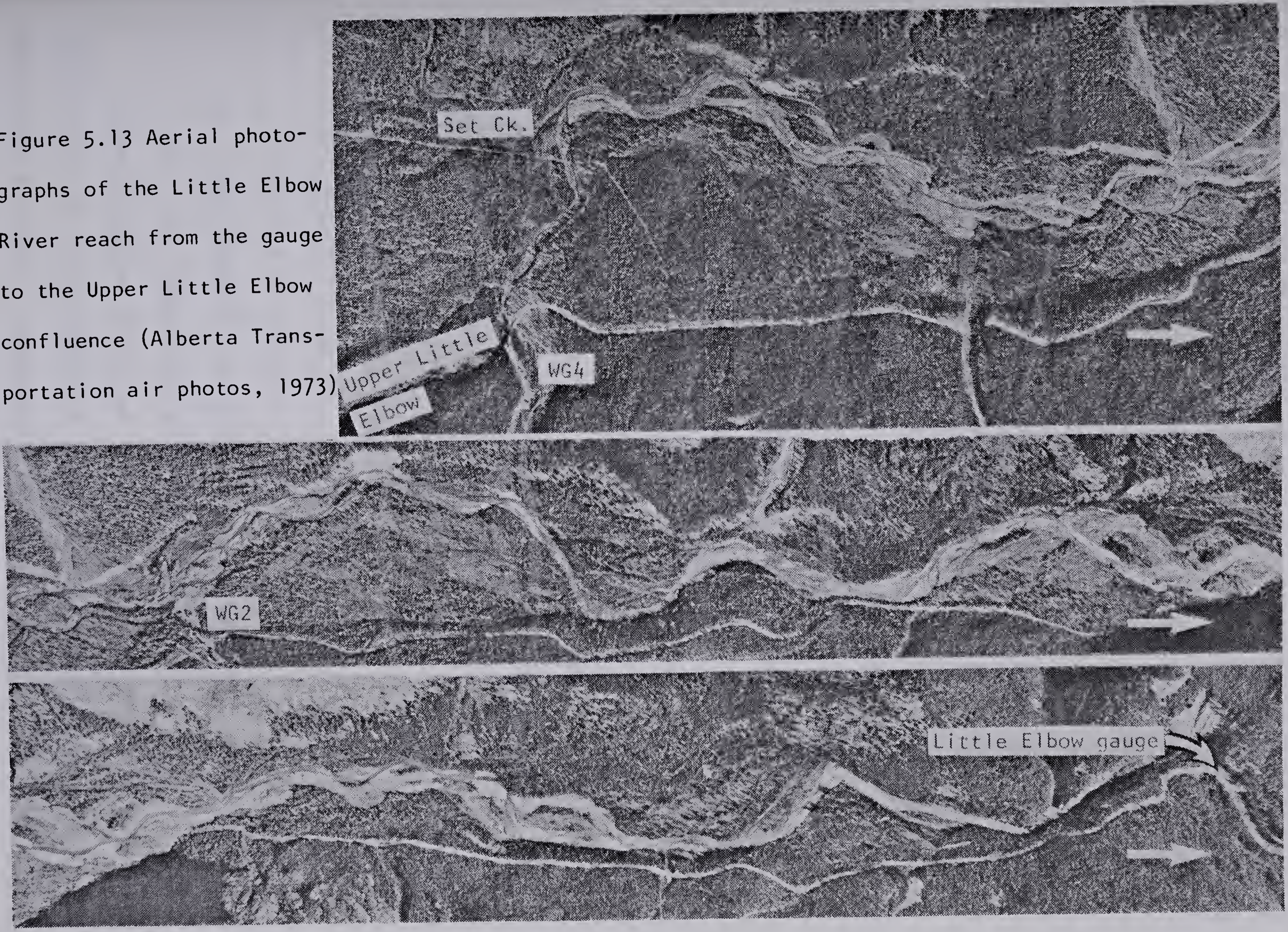

Figure 5.13 Aerial photographs of the Little Elbow River reach from the gauge to the Upper Little Elbow confluence (Alberta Transportation air photos, 1973)

relationship between discharge and the active width of transport is poorly defined (Figure 5.14), and the active width does not appear to be related to the grain size of the material in transit. Transport of bedload appears to be preferentially located in the shallower parts of the channel, not the thalweg (Figure 5.15). This situation is not unique to the Elbow River (Hudson, in press (b)). Jonys (1976) found that, in the Vedder River, most transport occurred along the north bank of the channel, over a submerged bar, and that very little movement occurred in the deepest part of the channel (Figure 5.16). Leopold and Emmett (1976) also noted movement as sheets or tongues during high flows over a permanent gravel bar in the left third of the channel of the East Fork River.

The variation in transport across the channel at Bragg gauge can be explained, by variations in grain size and shear stress across the channel , (Figure 5.15) using Shields' relation, and measured grain size:

$$T/(\gamma s - \gamma) = 0.03 \qquad \text{...(5.1)}$$

Where: T = shear stress (kg/m^2)

γs = specific weight of sediment (2650 kg/m^3)

γ = specific weight of water

D = representative grain size in meters

0.03 = Tc suggested by Neill (1968)

Most studies consider the average size of the largest particles as the competent size (Baker and Ritter, 1975). Kellerhals (1967) suggested the use of D_{90} to represent the

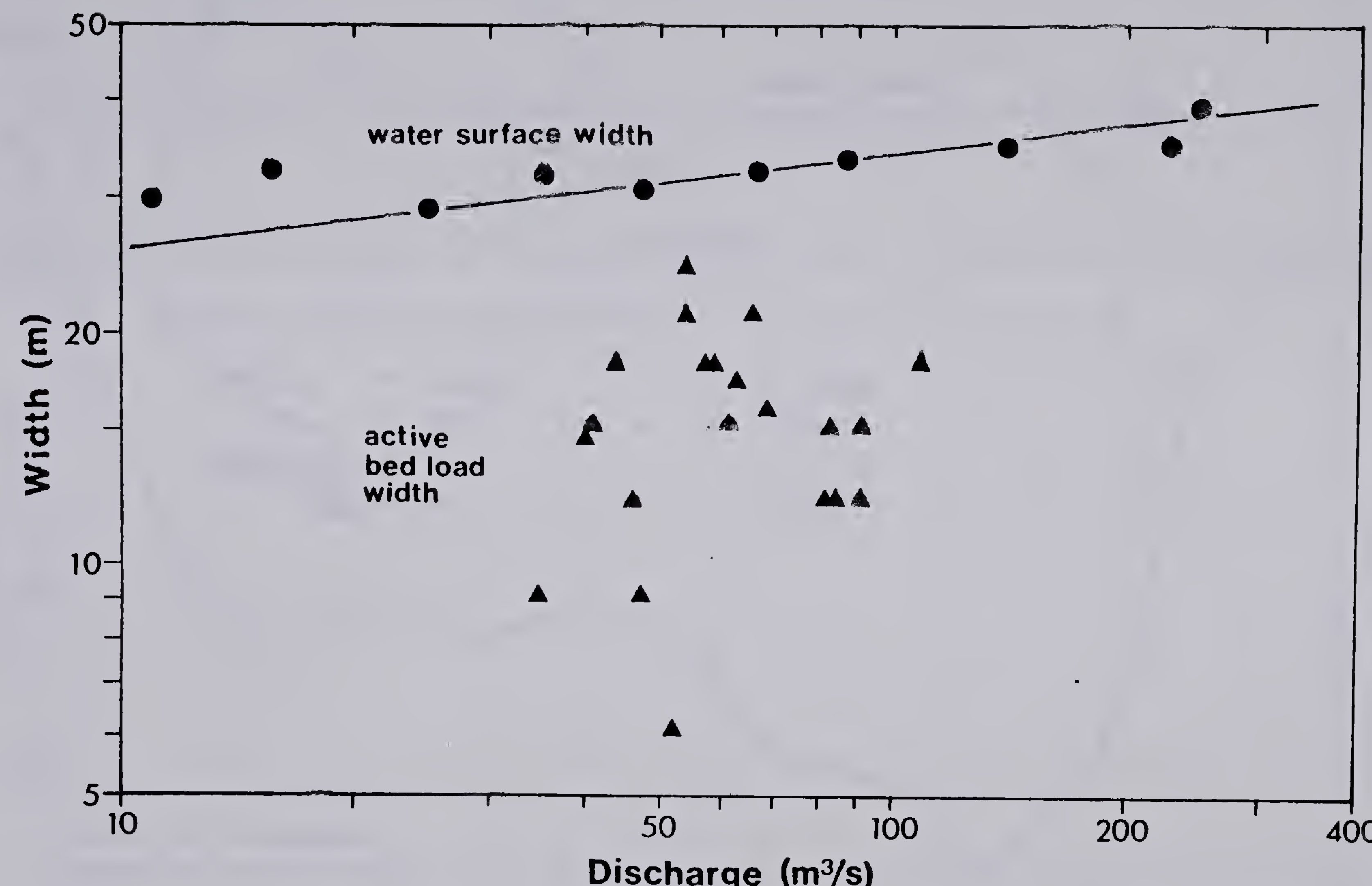

Figure 5.14 Water surface width, and measured width of bed load transport, against discharge, Elbow River at Bragg gauge

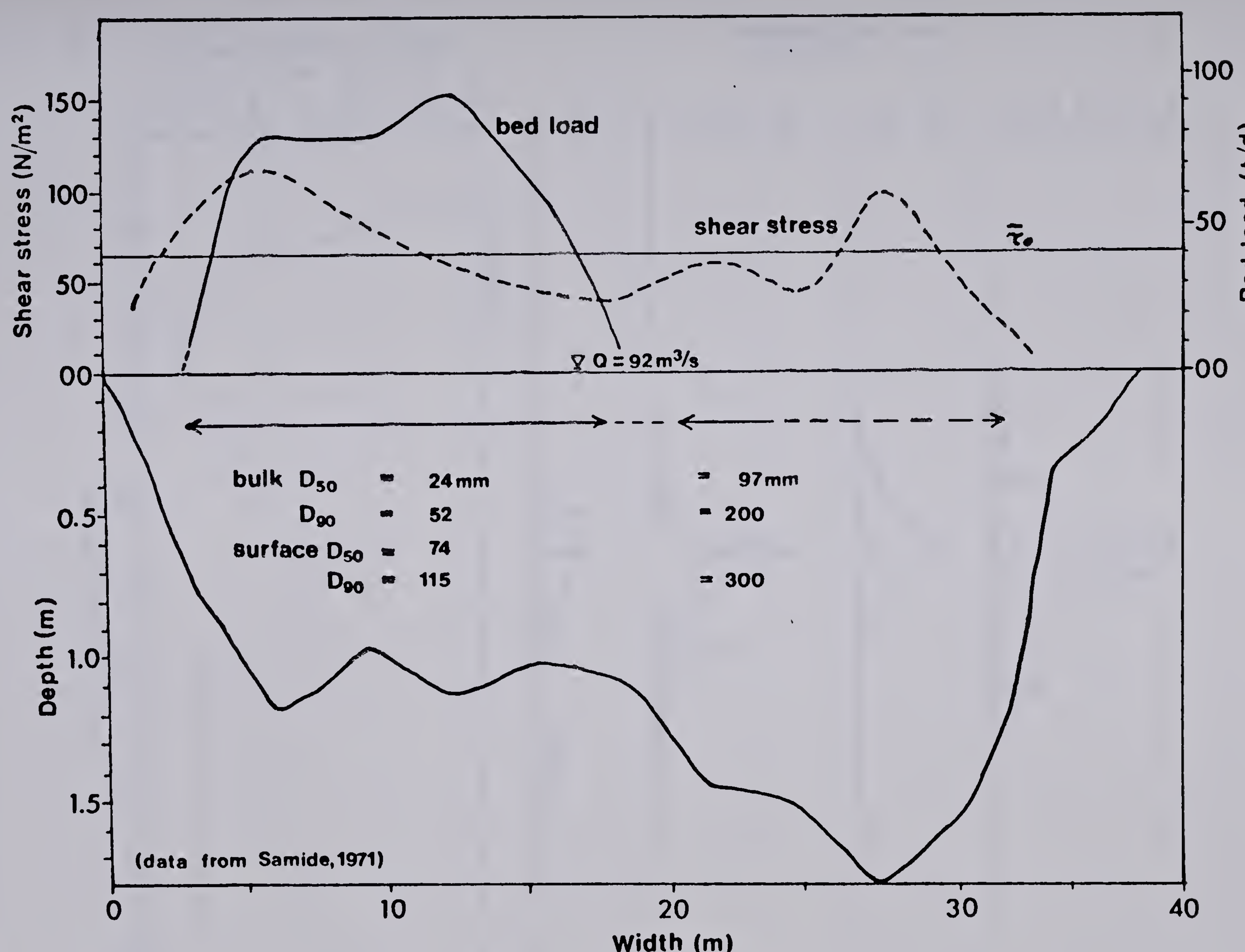

Figure 5.15 Cross - sectional variation in bedload and shear stress, Elbow River at Bragg gauge

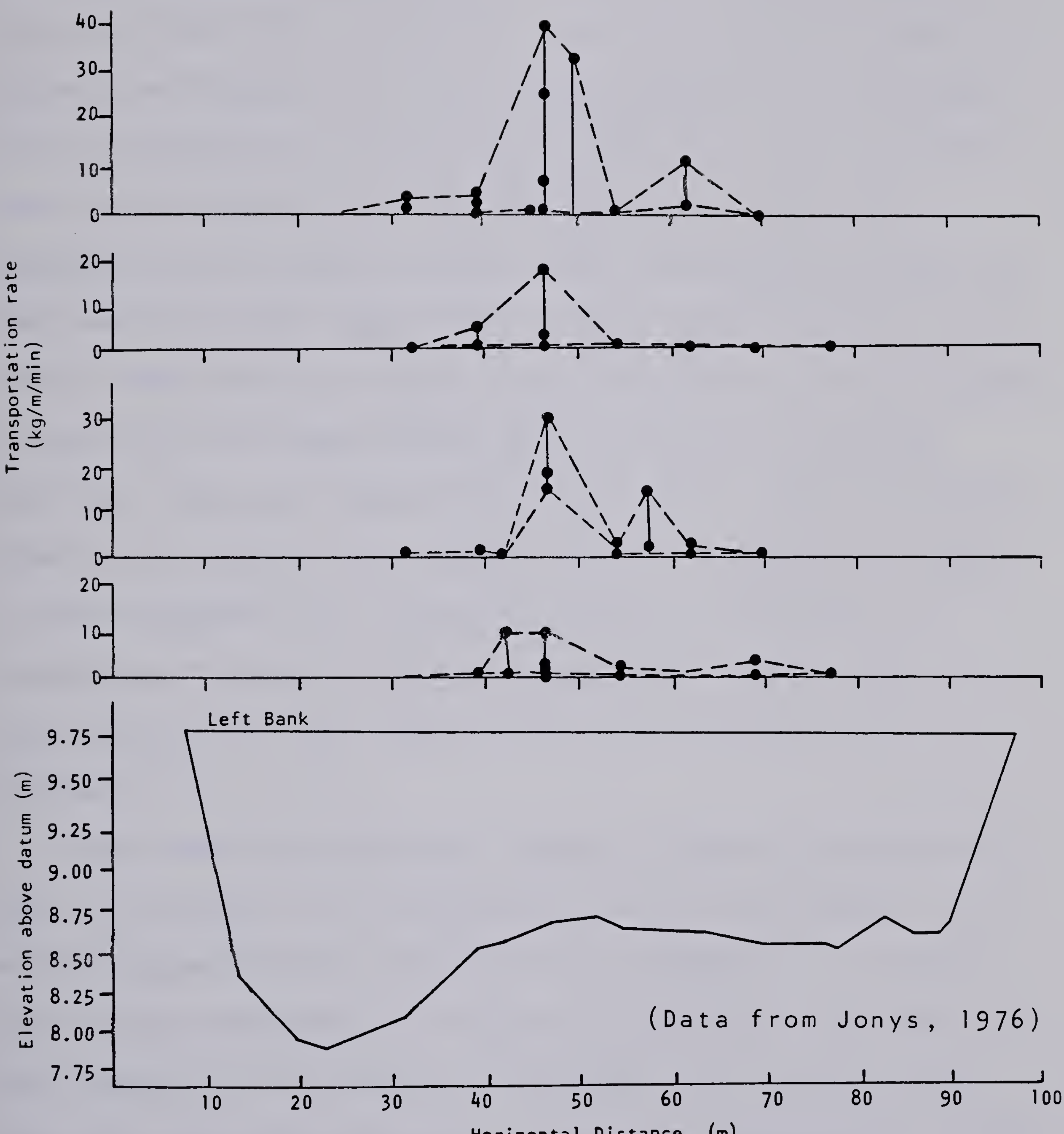

Figure 5.16 Cross - sectional variation in bedload, Vedder River near Yarrow, British Columbia

largest particles. If D_{90} is used, the stability analysis suggests that the average shear stress is competent to move the bar deposits (Tc = 55.8 N/m^2). However, the available shear stress is insufficient to move the thalweg deposit (To max = 60.8 ≤Tc = 146 N/m^2). In other words, the bar deposits are paved, but the thalweg deposits are armoured, until a discharge in the order of 500 m^3/s. This is an 85 year flood (Figure 3.5). The distinction is that the pavement can be moved at high flows, whereas the armour is the coarse surface layer which "never" moves (Parker et al., 1982). There have only been a few occasions, in the 72 years of record in the Elbow basin when a 500 m^3/s flood has occurred. The most recent major event was a 430 m^3/s flood immediately before Hollingshead began his sampling program in 1967 (Appendix 2). Particle packing, to produce an "underloose" deposit (Church and Gilbert, 1975), may increase the critical shear stress for initiation of movement.

The stability analysis, however, does not explain the lack of transport in the deepest part of the channel - it merely suggests that the flow is not competent to entrain the armoured deposits. In terms of competence, the flow in the thalweg is sufficient to transport any material which may spill over from the bars. Hydrophone measurements verify the bedload sampling (Samide, 1971; Jonys, 1976). Thus, the lack of transport in the deeper part of the channel is an anomaly, which is not explicable at present. Flume

experiments might reveal the paths of bedload transport in a cross - section and downstream.

Particle size variations

The mean size of particles in transit tends to decrease downstream, paralleling the downstream variation in river bar and river bank composition (Figure 2.7). The downstream variation in particle size is indicative of hydraulic sorting of particles, given that abrasion losses in size are expected to be small:

$$D = Do \exp (- ad\ x) \qquad ...(5.2)$$

where D is a characteristic particle diameter, at some distance x, measured downstream from some arbitrary starting point (x = 0 where D = Do), and ad is the coefficient of diminution (Church and Kellerhals, 1978).

Shaw and Kellerhals (1982) estimated the coefficient of diminution (ad) to be 0.003 km for limestone, in the nearby Red Deer River. Hence, as an extreme, a particle travelling from the headwaters to the Bow River would be expected to lose about 30% of its initial size due to in - transit and in - place abrasion. Losses would be smaller for the majority of particles travelling over shorter distances.

Particle size variations also occur over time at a given cross - section. In the Clearwater, East Fork, Snake and Vedder Rivers, and Oak Creek, two sediment discharge regimes have been recognised. At low discharges "throughput" of sand occurs over a paved gravel bed. At higher discharges

the pavement is breached, and all sub - pavement particles are moved. This situation appears to be the case in the Elbow River. The material involved in low discharge bedload transport, in the Elbow River, is coarser sand and very fine gravel (Figure 5.17). The bar pavement does not have appreciable amounts of this material, which tends to be deposited at the distal end of bars. Truncation of the bars and banks at relatively low flows provides the supply of these materials to the stream in addition to tributary inputs of finer materials. At higher discharges, the bar sub - pavement deposits and banks are eroded and, the material in transit is almost equal in size to these materials (Figure 5.18).

D. PREDICTION OF BED MATERIAL TRANSPORT

Introduction

The objective of this section is to find an appropriate means by which bedload transport can be predicted throughout the Elbow River system. There are several approaches to the prediction of bedload in sediment transport studies: The bedload can be ignored (e.g. McPherson, 1975), or arbitrarily defined as some fraction of known or estimated suspended sediment yield (e.g. ASCE, 1975). More directly, load can be calculated from sediment traps such as reservoirs, weirs or pits in the stream bed (e.g. Hollingshead, 1968; Nanson, 1974), or bedload may be measured at a

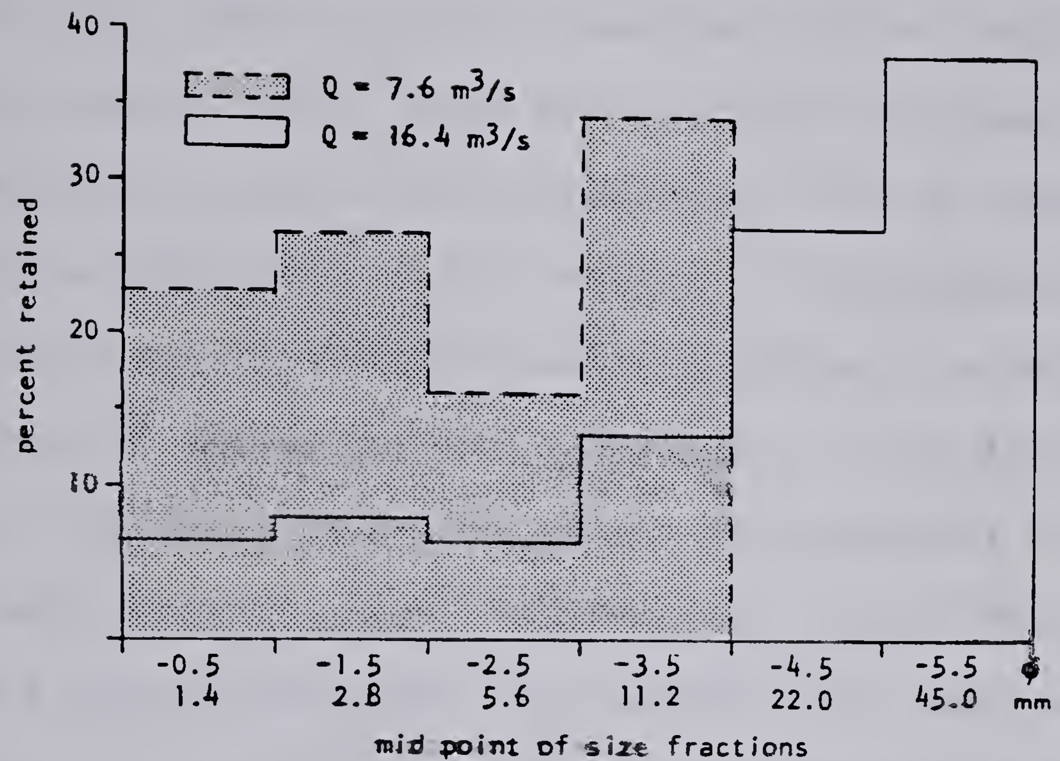

Figure V.17 Histogram of bed material fractions transported during high and low discharges, Little Elbow River gauge

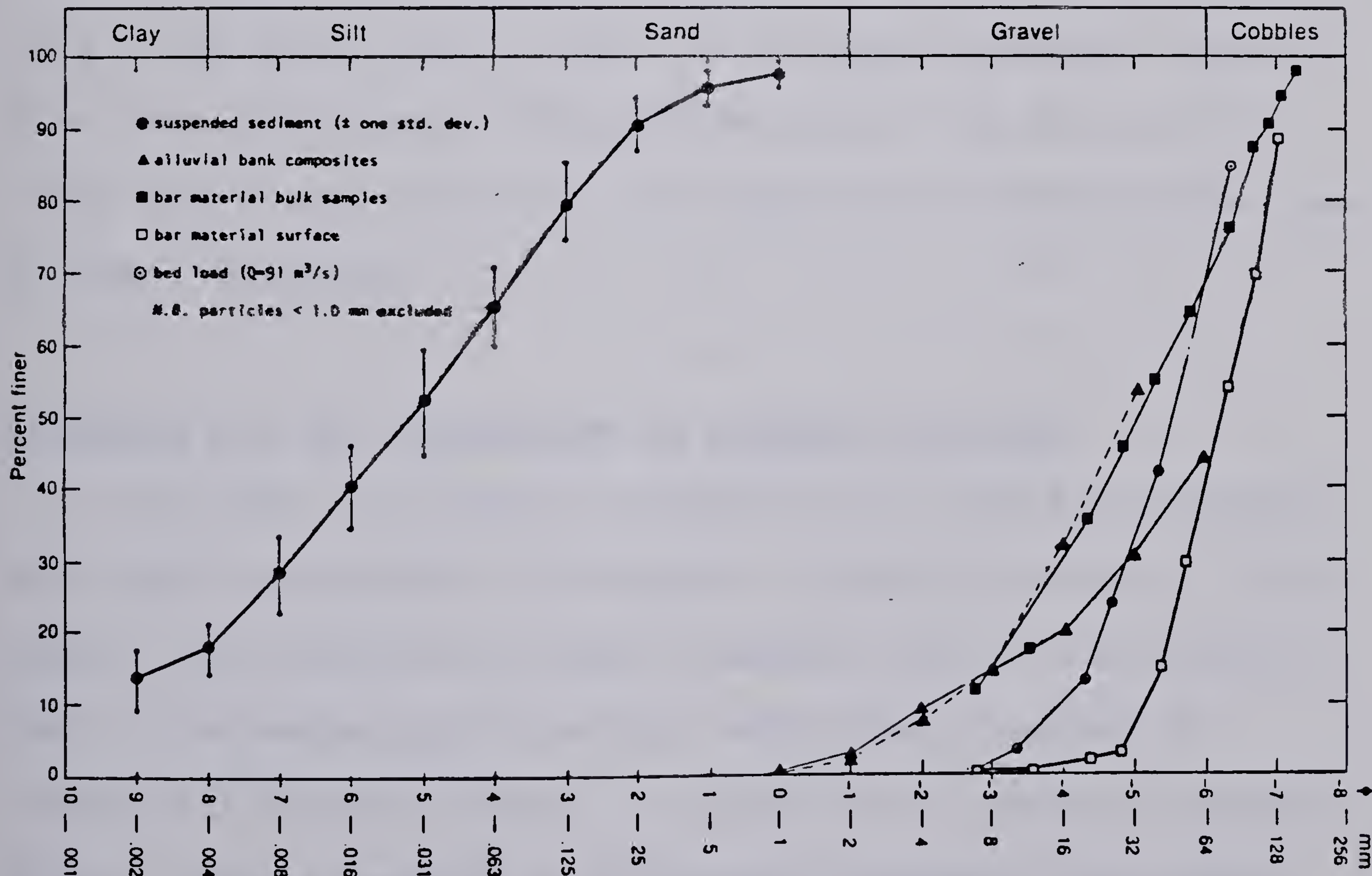

Figure 5.18 Composition of river bank, bar deposits, and sediment load in the Bragg gauge river reach

place in a stream to develop load - discharge relationships (e.g. Strand, 1975). Finally, bedload may be predicted by formulae (ASCE, 1975). Each approach has problems.

Bedload in gravel bed streams can not be ignored, nor assigned an arbitrary value relative to suspended sediment discharge, because the bedload is important in determining the hydraulic and geomorphic character of the channel. Further, little is known regarding the rates of movement in gravel bed rivers (Leopold and Emmett, 1976). Hence, there is little subjective basis for defining the bedload component as a fraction of the total load. Utilization of bedload traps, such as weirs, pits and reservoirs, often do not adequately define the time rate of transport. Until recently the use of bedload samplers had been restricted to only a few gravel bed rivers. As a result, bedload formulae have remained largely untested in gravel bed rivers. The Elbow River data provide a rare opportunity to evaluate some of these formulae.

Criteria for the evaluation of bedload formulae

Two types of temporal variability in bedload transport have been recognised: (a) supply - limited transport, which occurs over a period of days to weeks, and (b) variability due to the mechanics of motion, which has a period of seconds to minutes (Hudson, in press (a)). Sediment supply may be limited because of stream bed pavement development (Milhous and Klingeman, 1971), resulting in an "underloose"

boundary (Church and Gilbert, 1975). Perhaps more importantly, particularly in upland rivers, bedload transport may be largely controlled by river bank and riparian inputs into the stream channel (Nanson, 1974; Griffiths, 1980). Thus, in contrast to lowland rivers, bedload is not controlled solely by bed and hydraulic variables (Griffiths, 1980).

In addition to supply - constrained sediment transport, considerable variability in the rate of transport may occur, for given hydraulic conditions, over a period of seconds to minutes. These fluctuations have been attributed to the mechanics of motion in a previous discussion. Bedload samplers usually obtain a sample of the material in motion over a period of minutes. As a result the sample represents an integration of a fluctuating rate of motion. The sample obtained therefore represents some sort of time - averaged load, rather than an instantaneous load. Bedload formulae are designed to describe capacity loads. Academically, the capacity load would be the instantaneous maximum load. Practically, the capacity load would be the maximum load which is able to be sustained by a given flow. Because the instantaneous load fluctuates with time, the sustained load is represented by a time integrated average of the loads. This may be referred to as the mean capacity load. The duration of the bedload sampling would determine what this time - averaged load represents with respect to the actual mean capacity load. Hudson (in press (a)) suggests that the

Elbow River bedload data adequately represent the mean capacity load. Therefore, the relevant criteria for the evaluation of bedload formulae should be the ability of the formulae to describe the mean capacity loads, which are the data that represent the upper boundary, rather than a "best fit" of the data points.

Evaluation of bedload formulae

Eleven formulae have already been tested using Elbow River at Bragg gauge data (Hollingshead, 1968; Church, 1976; and McLean, 1980). However, much of the data used to test the formulae has since been revised. Load discharge relationships from each of the formulae were converted to metric units and plotted against the revised data. The formulae predictions are presented in Figure 5.19. In addition four formulae were evaluated by the writer at four sites on the Elbow River (Figures 5.20 to 5.23).

The original versions of each formula, and Hollingshead's (1968) bed material descriptions were used to estimate bedload for given discharges at Bragg gauge (Figure 5.19). It is apparent that only one formula, Colby and Hubbell's (1961) "Modified Einstein equation", is close to describing the measured rates of bedload transport at Bragg gauge. The next best formulae produce discharge - load relations which are of approximately the correct shape, but are in the wrong position, producing under - estimates of bedload (Einstein, 1950), or over - estimates of bedload

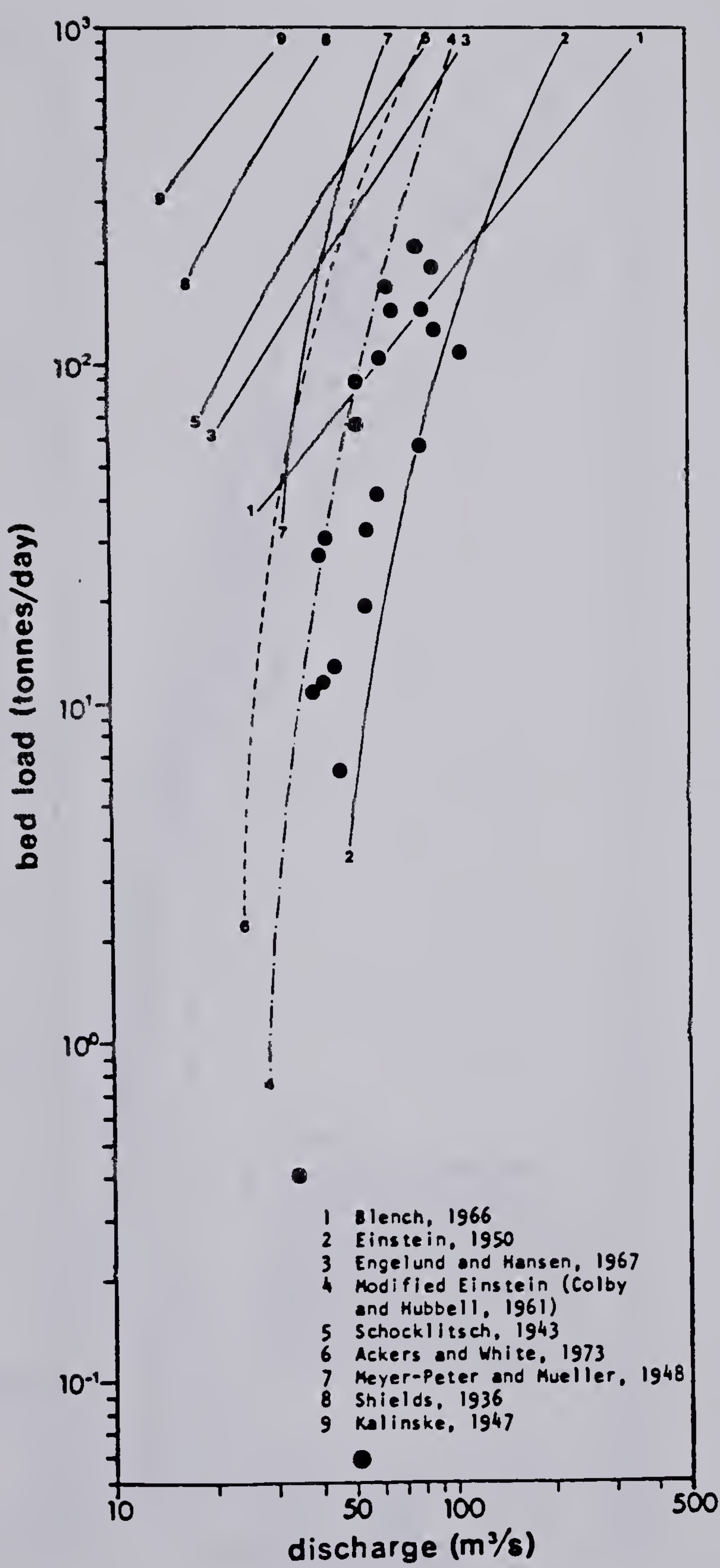

Figure 5.19 Evaluation of selected bedload formulae with revised Elbow River at Bragg gauge data

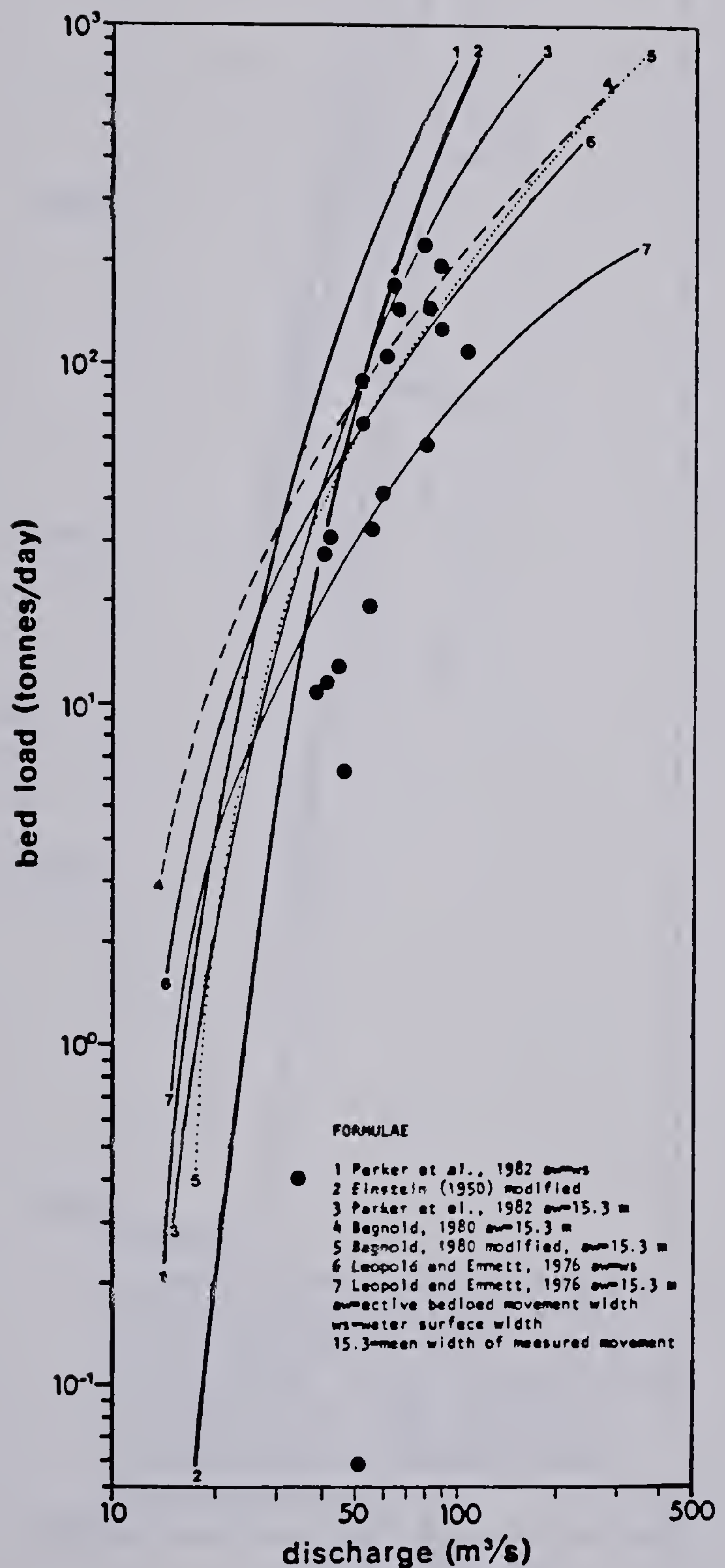

Figure 5.20 Evaluation of selected recent bedload formulae with revised Elbow River at Bragg gauge data

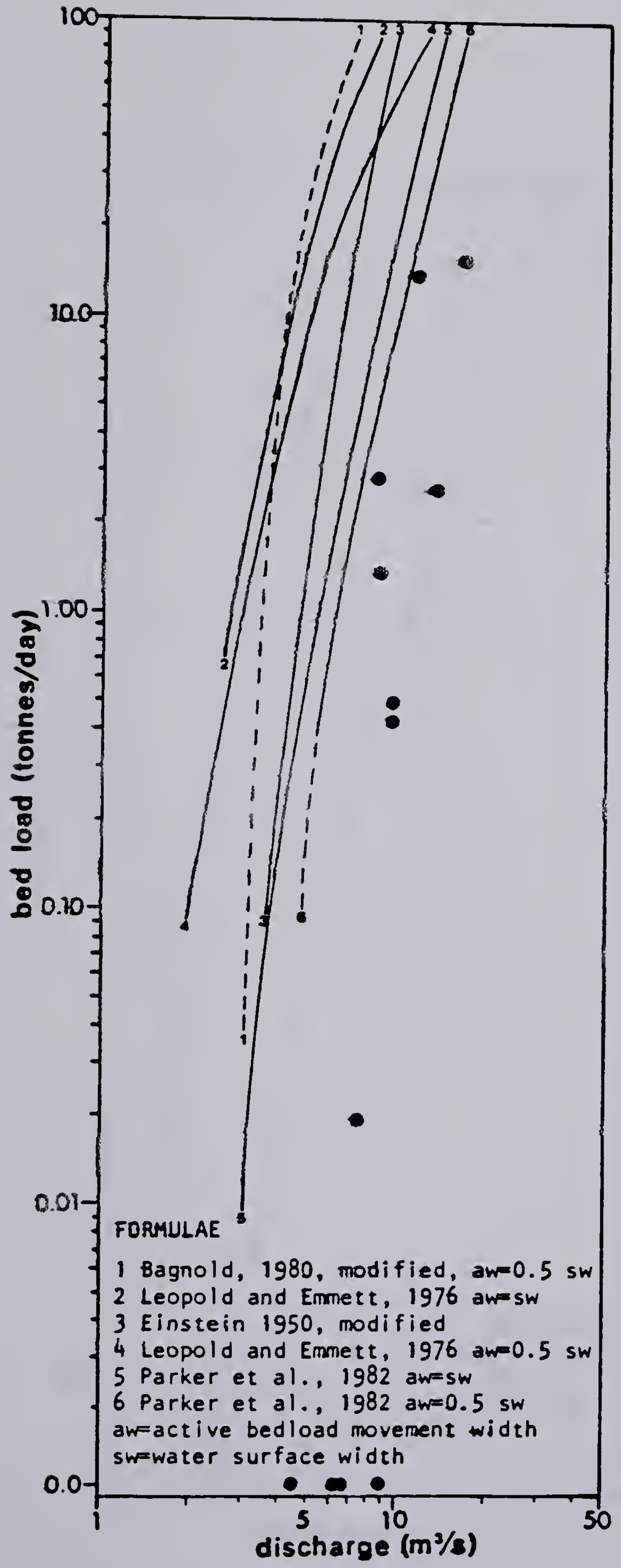

Figure 5.21 Evaluation of selected bedload formulae at Little Elbow gauge

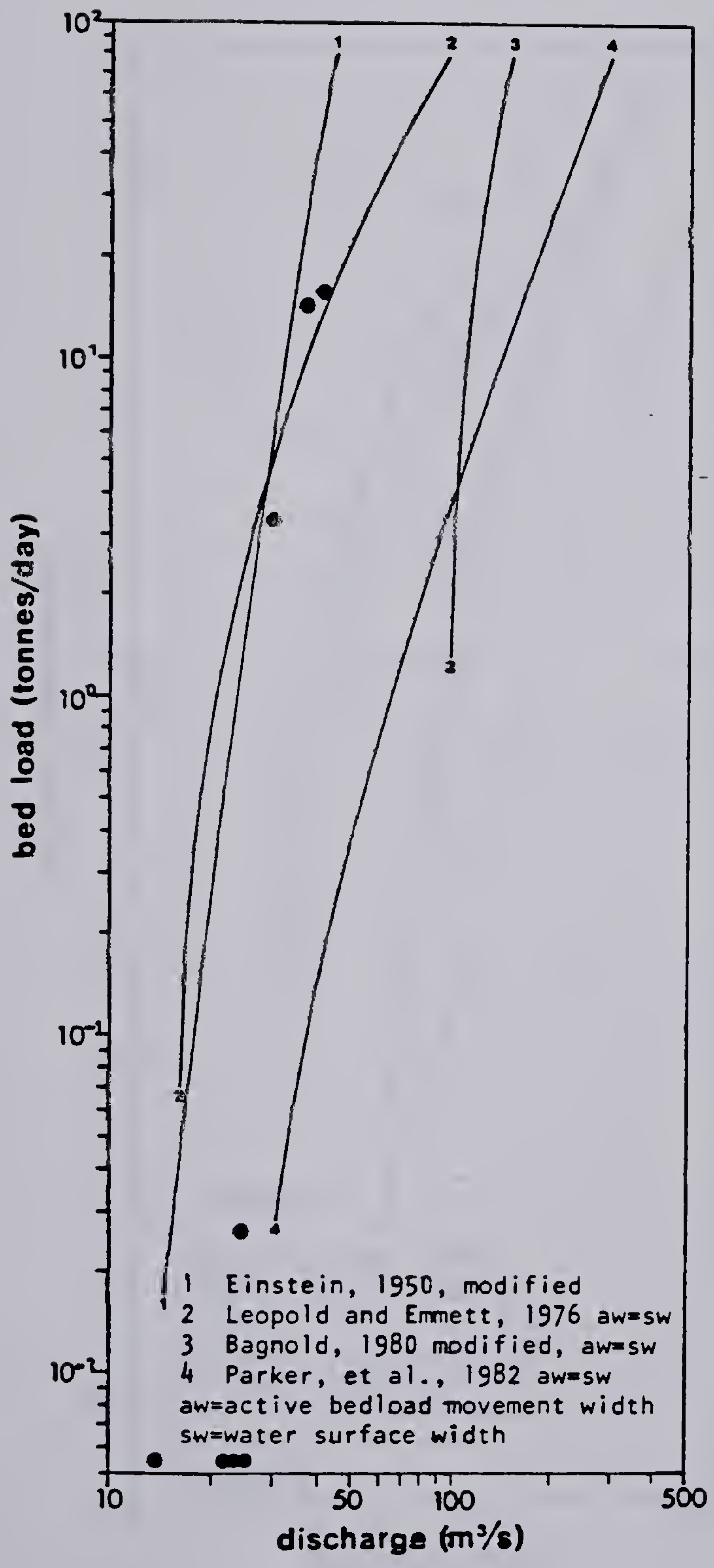

Figure 5.22 Evaluation of selected bedload formulae at Gardener gauge

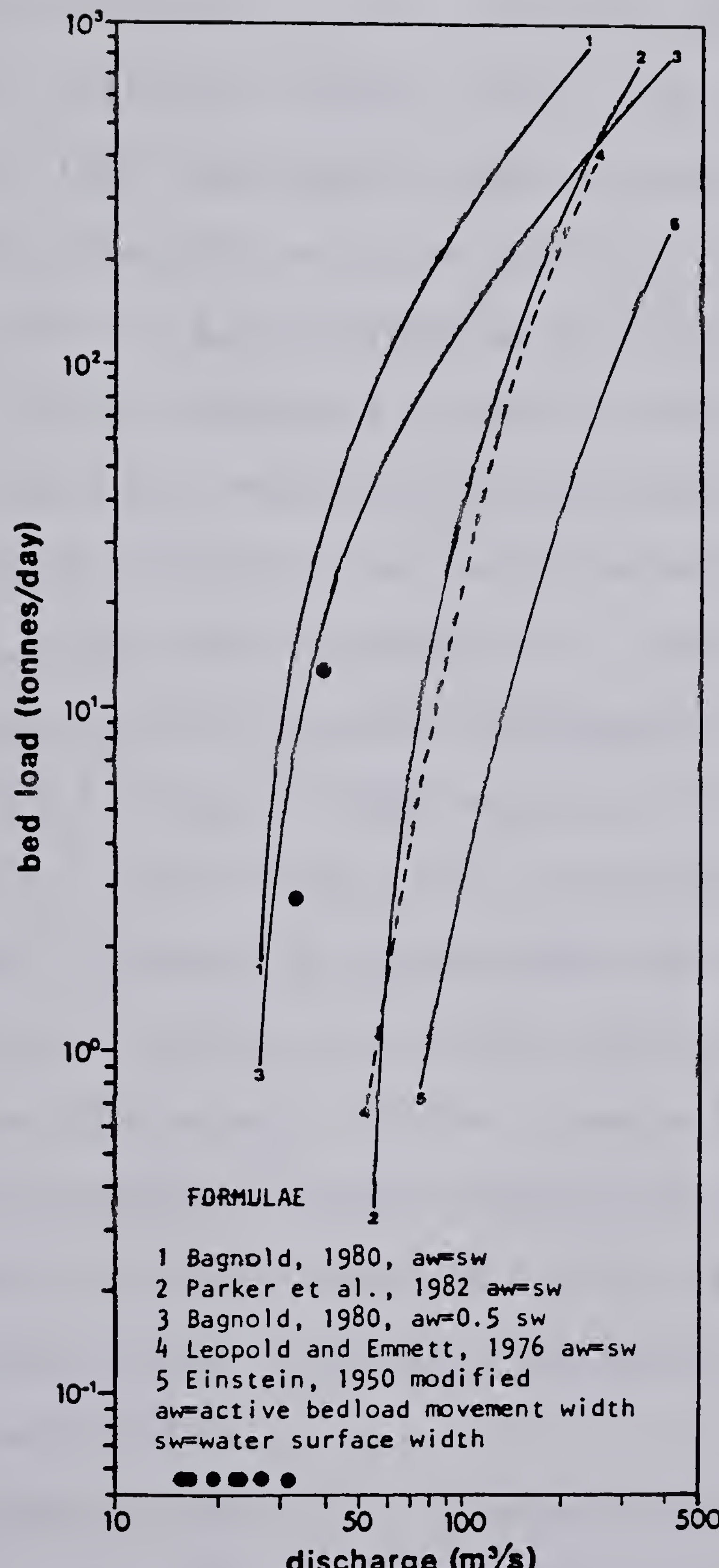

Figure 5.23 Evaluation of selected bedload formulae at Sarcee gauge

(Ackers and White, 1973; and Meyer - Peter and Muller, 1948). The remaining formulae, by Blench (1966), Engelund and Hansen (1967), Kalinske (1947), Schocklitsch (1943) and Shields (1936), which are considered to be reliable or popular in the literature (Church, 1976), produce bedload rating curves which are of the incorrect shape and plot in the incorrect position. Church (1976) also evaluated the Schocklitsch (1934) and Meyer - Peter equations. These plot close to the curves of the later versions of these formulae. Hence, they are not illustrated in Figure 5.19.

Church (1976) attempted to modify the parameters of the latter equations by incorporating measured width of bed movement, and by changing the roughness values and critical shear stress values where appropriate. However, the equations were still not able to adequately describe the measured bedload. McLean (1980) unsuccessfully tried to modify the Ackers and White (1973) and Meyer - Peter and Mueller (1948) formulae to better describe measured bedload on the Clearwater, Elbow, Snake and Vedder Rivers. He concluded that the Ackers - White formula does not appear applicable to gravel bed rivers, when mean hydraulic geometry data are used. The Meyer - Peter and Mueller (1948) formula tended to over - estimate the measured discharge by a factor of about five.

Hollingshead (1968:73) truncated the bulk bar material samples by excluding material less than 5.34 mm (1/4 inch) in his modified (Colby and Hubbell, 1961) Einstein analysis.

The sediment size data show that suspended sediment is almost invariably less than 1.0 mm in size, and that materials in the 1.0 to 6.34 mm range constitute a significant proportion of the bedload (Figure 5.18). When these finer materials are included in analysis, the lower discharge portion of the curve shifts downward, suggesting that transport of the fine materials takes place at low discharges (Figure 5.20). Einstein's (1950) bedload function has been computerized. His Figures 4, 5, 6 and 10 are approximated by a series of equations. Measured discharges, velocities and surface water width, are used in the calculations because, in the original version, resistance was predicted using the Einstein and Barbarossa (1952) relationship, which results in over - estimates of depth and under - estimates of mean velocity (McLean, 1980). Also, the hydraulic radius due to grain roughness was computed from the Manning - Strichler equation instead of the "bar resistance graph" proposed by Einstein. The results are almost identical to the graphical simplification of Einstein by Colby and Hubbell (1961).

The computer version of the Einstein formula was used to predict bed load at the other gauge sites, using mean hydraulic geometry and bulk bar samples, to define grain size (Figure 5.20 to 5.23). Summary statistics, of the hydraulics and the grain size distributions used in the calculations, are given in Table 5.6.

Table 5.6 Summary statistics of variables used in bed load prediction

Station	hydraulics					grain size (mm)								
	Slope m/m $\times 10^{-3}$	mean velocity m/s	mean depth m	surface width m	kinematic viscosity	d_{95}	d_{80}	d_{65}	d_{60}	d_{50}	d_{40}	d_{35}	d_{20}	d_{5}
Little Elbow gauge	14.50	$0.54\ Q^{0.49}$	$0.23\ Q^{0.33}$	$8.14\ Q^{0.18}$	1.40×10^{-6}	118	74.0	57.0	51.5	43.5	36.5	35.0	24.5	11.0
Falls gauge	8.11	$0.46\ Q^{0.39}$	$0.17\ Q^{0.35}$	$13.01\ Q^{0.26}$		100	50.0	35.0	31.0	25.0	19.0	16.0	6.8	1.8
Bragg gauge	7.45	$0.28\ Q^{0.48}$	$0.20\ Q^{0.37}$	$17.86\ Q^{0.14}$		120	56.0	38.0	34.0	27.0	20.0	17.5	9.0	3.4
Gardener gauge	7.07	$0.24\ Q^{0.55}$	$0.22\ Q^{0.22}$	$18.52\ Q^{0.22}$		120	56.0	38.0	34.0	27.0	20.0	17.5	9.0	3.4
Sarcee[1]	1.776	$0.45\ Q^{0.22}$	$0.13\ Q^{0.64}$	$16.73\ Q^{0.15}$		42.0	28.0	21.5	20.5	18.0	15.0	13.8	10.5	4.3
Sarcee[2]						34.0	20.0	14.0	12.5	10.4	8.4	7.5	5.0	2.2

Sarcee[1] bed material subpavement > 1.0 mm
Sarcee[2] bank material composite > 1.0 mm

The Einstein bedload function describes the measured bedload on the Elbow River at Bragg gauge and Gardener gauge reasonably well, when the bar deposits are used as the grain size input, and mean hydraulic geometry measurements are used (Figure 5.20 and 5.22). However, there are problems with the approach which makes prediction of loads at unmeasured locations uncertain. The equation predicts bed-load from the total stream width, whereas it is known that approximately half of the stream channel is armoured. This means that the equation is under - predicting load in the active width and predicting movement where it does not occur. If the actual active width increases or decreases, relative to the total width, then load would be under - predicted or over - predicted, respectively. A further constraint on the Modified Einstein bedload function is the range of applicability. Bedload is under - predicted by a factor of about 10 at Sarcee gauge, and over - predicted by a factor of about 10 at Little Elbow gauge.

Leopold and Emmett (1976) express unit bedload, as a function of unit stream power (W) and D_{50} of the sediment, in a nomograph (Figure 5.24). Unit stream power is derived from reach hydraulic geometry. Measured bedload at Bragg gauge is not well described by the nomograph, using the bulk bar deposit D_{50}. Although the nomograph is for bedload capacity, load is under - predicted at high discharges, even if the active width of gravel bed movement is assumed to equal the surface water width of the stream. Further, the

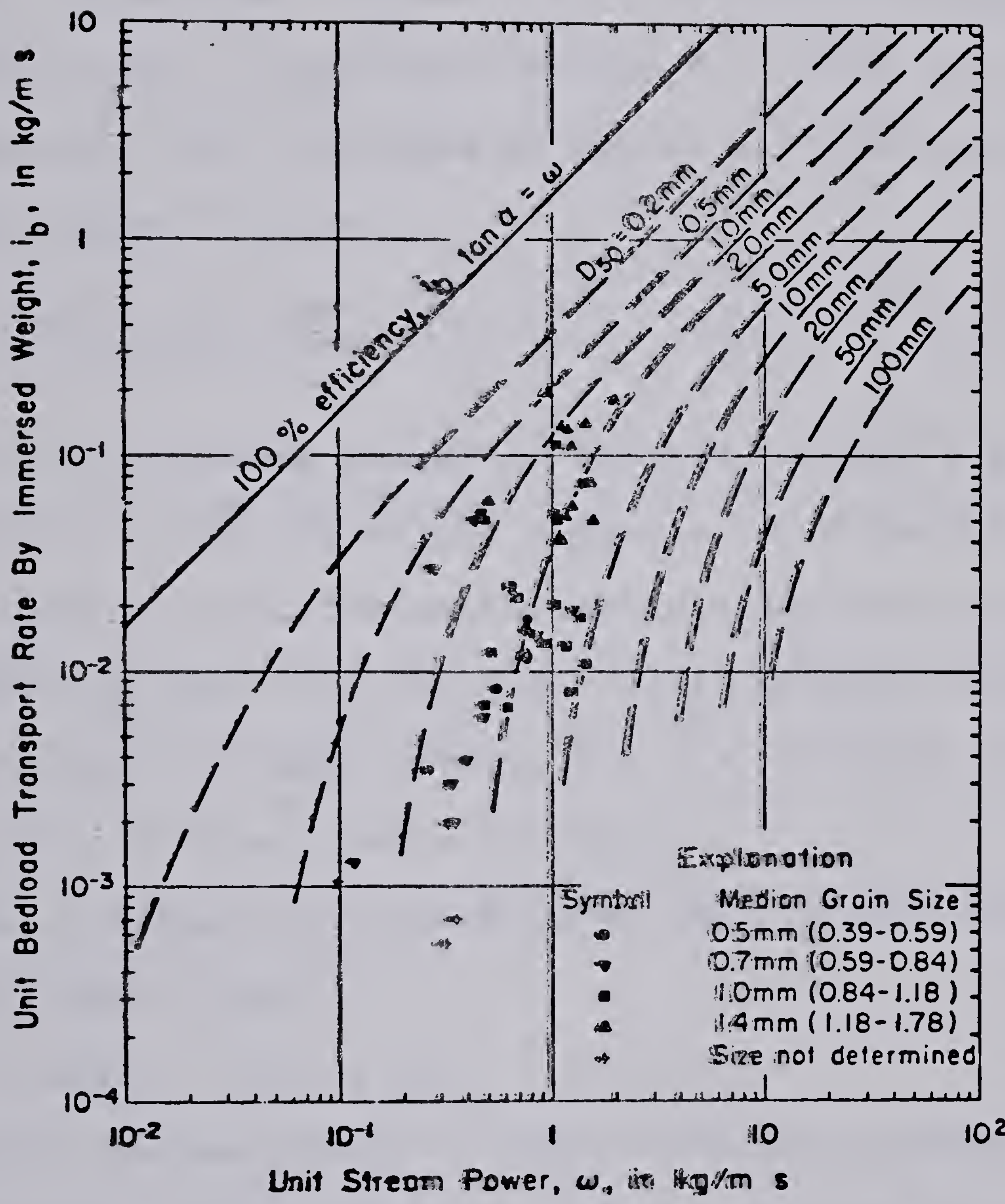

Figure 5.24 Leopold and Emmett's 1976 nomograph of unit bed-load discharge against unit stream power

threshold discharge is under - predicted at Bragg gauge (Figure 5.20). The Leopold and Emmett (1976) approach fits the limited Gardener gauge data reasonably well (Figure 5.22). At Little Elbow gauge bedload is grossly over - predicted at high discharge, and the threshold of movement is under - predicted (Figure 5.21). At Sarcee gauge bedload is grossly under - predicted (Figure 5.21 to 5.23).

Bagnold (1980) proposed an empirical correlation of bedload transport rates:

$$ib = [(W-W_o)/(W-W_{o*})]^{3/2} \cdot (y/y_*)^{-2/3} \cdot (D/D_*)^{-1/2} \cdot (ib_*) \quad \ldots(5.3)$$

Where: ib is immersed weight of bedload transport per unit width (ib x 1.61 = dry weight = kg/m - s); W is unit stream power (kg/m - s); W_0 is the threshold value of W at which D_{50} starts to move; D = D_{50} grain size; starred values are constants from Williams (1970): W - W_0* = 0.5; Y* = 0.1; D* = 1.10 x 10^{-3}; all units in kg.m.s.

The critical unit stream power (W_0) is derived by Bagnold (1980) from:

$$W_0 = 290\ D^{1.5}\ \lg(12\ Y/D) \quad \ldots(5.4)$$

where Y is the mean depth of flow in meters. A depth value of 1.035m was used by Bagnold (1980), hence W_0 was 3.1. An alternative approach was followed to estimate a representative mean depth of flow, Y. Shields equation (eqn. 5.1) was used to predict the threshold shear stress for D_{90} of the bar deposit. Hydraulic geometry relationships were then used to calculate the corresponding discharge, hence, unit stream

power.

Using Bagnold's (1980) approach for predicting critical unit stream power results in the over - prediction of the bedload transport, and an under - prediction of the threshold discharge to initiate motion of the bed material at Bragg gauge. The predictive power of the equation is improved if W_0 is derived from hydraulic geometry, Shields equation, and D_{90} of the bar deposits. The best fit of the Bagnold (1980) equation occurs at Bragg gauge, when the average measured width of bedload transport (15.3 m) is used with the alternative method of predicting W_0. This curve is closely approximated by using half of the measured surface water width to represent the active width of bedload movement.

Parker et al., (1982) developed equations which, they state "should be reasonable for small to medium - sized paved gravel - bed streams with steep (Elbow and Oak Creek) or moderate (Vedder River) slopes, which are not dominated by throughput sand bedload.":

for $0.95 < \phi 50 < 1.65$

$$W^* = 0.0025 \exp[14.2\ (\phi 50-1) - 9.28\ (\phi 50-1)^2] \qquad \ldots(5.4)$$

and for $\phi 50 > 1.65$

$$W^* = 11.2\ (1-(0.822/\phi 50))^{4.5} \qquad \ldots(5.5)$$

where: $\phi 50 = \tau^*_{50}/0.0876$

$\tau^*50 = \tau/\rho\ Rg\ D_{50}$ (=Shields stress)

$W^* = R\ q_B/[\sqrt{9.81}\ (H.S)^{1.5}]$ (= dimensionless total bedload in $m^3/m\text{-}s$)

R = specific weight of sediment (1.64)

q_B = volumetric total bedload per unit width of gravel bed

ρ = density of water

g = acceleration due to gravity

The Parker et al., (1982) formulae tend to slightly under - estimate threshold conditions, as does the Modified Einstein equation, and to over - estimate bedload at Bragg gauge. A better estimate of the rate of bedload transport is made when the width is adjusted to coincide with the mean measured width of bedload movement (Figure 5.20). The equations define bedload transport well at Little Elbow gauge, when transport is assumed to occur over half of the surface water width. At Little Elbow gauge, the range of application, $0.95 \leq Ø50$, may be extended to threshold conditions, where $Ø50 = 0.77$. However, at Gardener gauge, and Sarcee gauge, the equations grossly under - predict bed-load transport. In the case of Sarcee gauge, the formula is beyond the range of application ($0.95 \leq Ø50$) for discharges less than 40 m^3/s. At this point the measured loads far exceed the predicted load (Figure 5.23). However, this load may represent "throughput" load, perhaps from bank erosion. The Gardener gauge site may not adequately represent the variations in channel characteristics in this area.

E. PREDICTION OF BEDLOAD IN THE ELBOW RIVER BASIN

Introduction

Hollingshead (1968) used a flow duration curve and an equation relating bed load to discharge, to estimate annual bedload at Bragg gauge. To examine variations in transport over a long period, bedload was calculated from formulae

using mean daily discharges. It should be emphasised that the equations chosen to represent bedload - discharge relations quantify the mean capacity load rather than the best fit relation. In other words, the bedload rating curves describe the mean maximum measured rates of transport, rather than an average rate for a given discharge.

To calculate bedload transport from the selected equations, the bedload rating curves were divided into near linear segments, which were described by log - log regression models. For example, at Bragg gauge the Einstein curve (Figure 5.20) was split into 5 segments: 20 - 40, 40 - 80, 80 - 100, 100 - 200 and 200 - 300 m^3/s, which were described by linear models. A discharge of 20 m^3/s represents threshold conditions..

Model selection

The Parker et al., (1982) equations were chosen to represent bedload transport at Little Elbow gauge. Active width of bedload movement was assumed to be half of the surface water width, based on measured width of bedload movement, and D_{50} of the upstream bar deposits was used to represent grain size (Table 5.6, Figure 5.21 (Hudson, in press (b)). Annual bedload estimates are presented in Table 5.7. For the period before 1978, discharge at Little Elbow River gauge was predicted from relationships developed in Chapter 3 and Appendix 2. Bedload transport was calculated from these daily discharge estimates.

Table 5.7 Estimates of annual bed load transport, Elbow River basin hydrometric stations (tonnes)

Station/year	1981	1980	1979	1978	1977	1976	1975	1974	1973	1972	1971	1970	1969	1968	1967	Total	Mean
Little Elbow g.	1694	461	42	321	2	52	415	1629	244	629	564	812	1644	176	3065	11750	783
Falls gauge	5552	1240	66	471	1.4	73	935	4985	488	1554	1361	2125	5042	327	10807	35027	2335
Bragg gauge	23104	2994	90	760	0	187	1092	7521	2078	3816	10358	11326	40155	1254	97060	201795	13453
Sarcee gauge	1409	310	31	154	0	88	227	691	256	320	1086	1717	4017	173	4718	15197	1013

N.B.: Little Elbow gauge 1967-1977 estimated from Falls gauge ($LE = 1.552\ Fg^{0.817}$ $r^2 = 0.98$)
Gardener gauge bed load equal to Bragg gauge bed load

The Parker et al., (1982) formulae were also used to estimate bedload at Falls gauge because the approach, using D_{50}, and half of the surface water width to represent the width of bedload movement, reasonably describes the bedload measurements further upstream at Little Elbow gauge, and downstream at Bragg gauge (Tables 5.6 and 5.7). The bed at Falls gauge is also armoured, or has exposed bedrock, for approximately half the channel width.

Estimates of bedload at Bragg gauge were made using the Einstein equation, with measured hydraulic geometry relationships, and the grain size distribution of material larger than 1.0 mm, from bulk bar deposits (Figure 5.21 and Table 5.6). Material less than 1.0 mm in size is transported as suspended sediment load (Figure 5.18).

The bedload data from Gardener gauge are well described by the Bragg gauge bedload rating curve. Because the discharges of these two stations are closely related, being only 12 km apart with no major tributaries in between (Appendix 2), the bedload capacity at Gardener gauge was taken to be similar to that at Bragg gauge.

Bagnold's (1980) equation describes the bedload capacity at Sarcee gauge when stream power is calculated from the Shields parameter and hydraulic geometry relationships, and if half the bed is considered to be active (Tables 5.6 and 5.7; Figure 5.23).

<u>Load estimates</u>

Estimates of the annual bedload transport, for the 15 year period 1967 to 1981, are summarized in Table 5.7. The estimates show that bedload transport capacity increases greatly downstream from the mountains through the foothills, but decreases dramatically in the plains zone. The average annual bedload capacity of the Elbow River basin above Falls gauge is about 17% of the load capacity at Bragg gauge, in the lower foothills (2335 vs 13453 t/yr). The capacity load at Sarcee gauge, in the plains above Glenmore Reservoir, is about 8% (1000 t) of the load in the mid foothills.

The average annual bedload is approximately one quarter of the capacity load at Bragg gauge (3325 against 13453 t). The average annual load was calculated from daily discharges, using a curve of approximately the same shape as the Einstein curve, which was drawn through the middle of the data, with an origin at $25 m^3/s$. Hollingshead (1968) drew a similar curve and, using a flow duration analysis, calculated an average annual bedload of 4000 t at Bragg gauge. At Little Elbow gauge the average annual bedload averages 21% of the capacity load for the period 1978 to 1981, when a best fit line is drawn parallel to the Parker et al., (1982) bedload rating curve, roughly through the middle of the data, with the origin at $6 m^3/s$. If the average bedload is approximately one quarter of the bedload capacity, then the average annual loads are 164 t at Little Elbow gauge, 580 t at Falls gauge, 3325 t at Bragg gauge and

250 t at Sarcee gauge.

F. DISCUSSION

There is considerable variation in bedload transport at various scales in time and space. Cross - sectional variations in bedload transport have been discussed previously, as have variations in time from minute to minute, at a point in a cross - section. The discussion here is limited to the macro - scale load variations downstream, and at a point over time.

Bedload in the Elbow River system appears to be derived from a few mountain tributaries, limited riparian erosion, and general channel erosion. As well as being a sediment source, the channel may act as a sediment sink. Low river banks suggest that major aggradation has taken place in the channel at the edge of the mountains, downstream, almost to Falls gauge (km 87 to 106). In this reach the river valley is broad, alluvium - filled, and the river flows in a braided pattern at higher flows. Further upstream, in the mountains, the valley slope is approximately twice as great as in the braided reach (0.024 against 0.013 m/m), and the river is confined in a steep sided, V shaped, valley.

In the river reach from Bragg gauge to Falls gauge (km 60 to 85) there is a tendency toward degradation, resulting in the exposure of bedrock steps in the stream bed (Seagel, 1971). Between Bragg gauge and Glenmore Reservoir there are three major areas of deposition. The river long profile,

from Gardener gauge upstream, shows a distinct "hump" in the stream bed. This hump starts at km 51.5 and extends downstream several km from Gardener gauge. River banks in this reach are relatively low. In one section of this reach, the river bed has aggraded over 1.5 m, so that gravel deposition is occurring over the floodplain. Upstream and downstream of the zone of aggradation, the cross - sections are similar to the degraded reaches further upstream (Figure 5.25). More recently, significant aggradation has occurred in the Twin Bridge area (km 29). The average river bed elevation at Twin Bridges increased by 0.60 m in the period August, 1972, to June, 1981. Further upstream and downstream bank height surveys suggest the channel is either stable, or may be degrading.

The areas of aggradation in the Elbow River between Bragg gauge and Glenmore Reservoir may represent "channel sediment waves". Griffiths (1979), working in the gravel bed Waimakariri River, New Zealand, describes channel sediment waves downstream of degraded reaches, and in aggrading reaches which have bank erosion and mass movements of material into the stream channel. "Evidently production of the waves... is a function of conditions within the geomorphologic regime of the catchment rather than the fluid dynamic regime of the river." (Griffiths, 1979:23). In the case of the Gardener aggadation zone, it appears as if the sediment wave developed in two phases. First, aggradation occurred over several km in a reach with a relatively

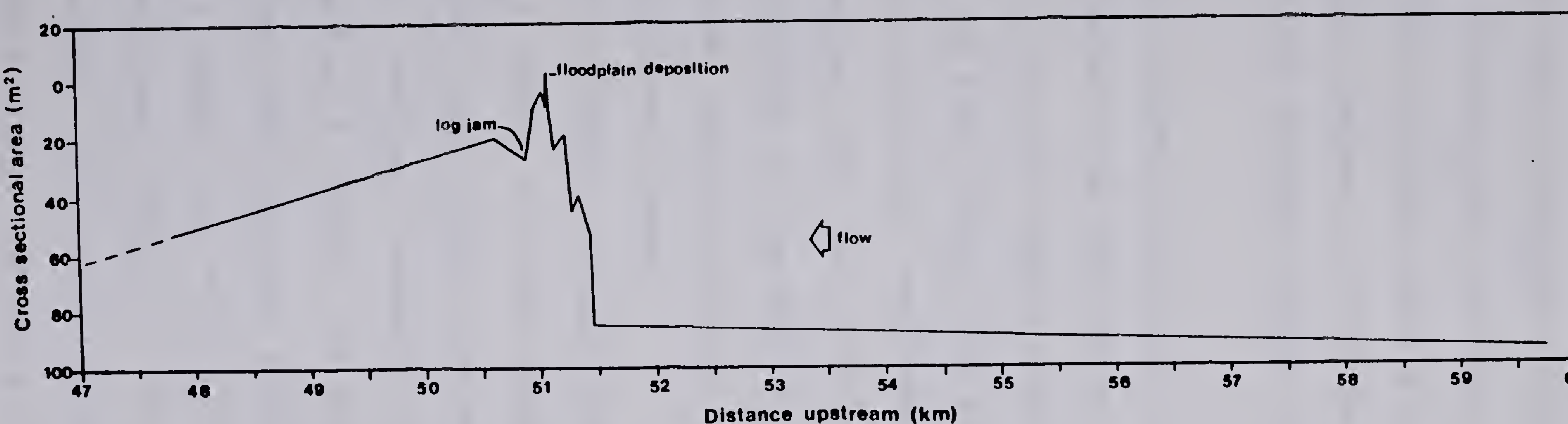

Figure 5.25 Cross - sectional area of the river channel in the lower foothills aggradation zone

uniform slope (Figure 2.2). Secondly, a more recent, and probably rapid build - up, occurred behind a log - jam in the aggrading reach. Subsequent deposition upstream of the jam may have occurred as a traction clog deposit (Moss, 1972). The 1950 air photos show that the river flowed along the north bank, whereas in 1972 the channel was partially entrenched in its present course, along the south bank 300m away.

The total volume of material involved in the bedload plug is estimated to be 185,120 m^3, which is roughly equivalent to the mean capacity bedload transport, at Bragg gauge, for a 20 year period. The air photos suggest that the majority of this material was deposited since 1972. Thus, sediment transport must vary significantly downstream in response to changes in the hydraulic regime and as a function of the relative position of macro forms (sediment waves) in the river channel.

Bedload is highly variable from year to year, at a given station (Table 5.7). For example, at Bragg gauge annual capacity load varies from zero (1977) to over 97000 tonnes (1967). Bedload transport is highly discharge - dependent, and as a result, the majority of transport takes place in a very short period, because the hydrograph is typified by a rapid rise and fall in discharge, at the time of the annual flood (Chapter 3 and Appendix 2). On average almost 90% of the total annual bedload is transported within 10 days (Figure 5.26). According to the capacity formulae

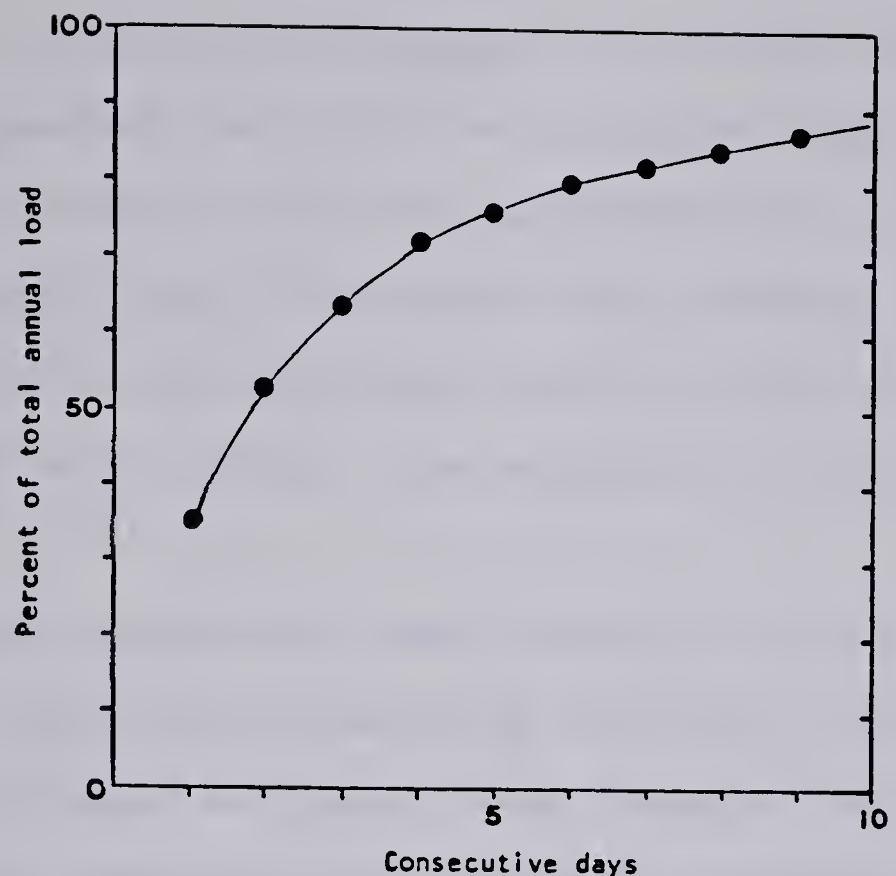

Figure 5.26 Cumulative percent of total annual bedload transported over time at Bragg gauge

chosen to represent each gauge site, bedload movement occurs on average about 50 days per year at Little Elbow and Falls gauge, 27 days per year at Bragg gauge and 20 days per year at Sarcee gauge.

G. CONCLUSION

Gravel bedload transport was measured, using a Helley - Smith bedload sampler, at several sites in the Elbow River system, during 1978 and 1979. In addition, large basket samplers were used to measure bedload at Bragg gauge in 1967, 1968 and 1969 (Hollingshead, 1968; Samide, 1971). The latter data have been revised.

New criteria are proposed to evaluate bedload formulae. It is suggested that formulae should describe the maximum rate of transport which is sustainable over time. This, the mean capacity load, recognises that sediment transport is controlled by sediment supply and by momentary extremes in transport which reflect the mechanics of bed material motion.

Eleven frequently used formulae, and four more recent formulae, were evaluated using the Elbow River bedload data and the proposed criteria. Some formulae were totally unsuitable, whereas others could be modified to adequately describe the measured bedload. However, no formula is able to accurately describe the measured bedload at all points in the system.

The Einstein (1950) bedload function, which was modified to include measured hydraulic characteristics, described capacity load at Bragg gauge and Gardener gauge very well (water surface slope 0.00745 and 0.00707 respectively and D_{50} 27 mm,). However, in the very steep Little Elbow River (slope 0.0145, D_{50} 43.5 mm) the equation over - predicted bedload by a factor of 10, whereas at Sarcee gauge (slope 0.001776, D_{50} 18 mm) the equation under - predicted measured bedload by a factor of 10. At Bragg gauge the bedload rating curves are identical to those calculated from Colby and Hubbell's (1961) "Modified Einstein" procedure, when the same grain size distribution is used.

The equations of Parker et al., (1982) also described capacity load at Bragg gauge quite well, when the active width of transport was taken to occur on the non - armoured portion of the bed. Their formulae provided the best description of mean capacity bedload at Little Elbow gauge. The Sarcee gauge reach is out of the range of applicability of the formulae, until very high discharges, when only small quantities of bedload transport are predicted.

The capacity nomograph of Leopold and Emmett (1976) under - predicts measured rates of bedload transport at each site, apart from Little Elbow gauge, where load is over - predicted. The Bagnold (1980) equation is also based on stream power. The formula was modified so that critical stream power was determined from Shields' equation, and hydraulic geometry, and the width of bedload movement was assumed to equal half the surface water width. These modifications describe the bedload at Bragg gauge better than the original equation. The equation adequately describes bedload transport at Sarcee gauge. Elsewhere in the system the formula over - predicts (Little Elbow) or under - predicts bedload.

The equations of Shields (1936), Schocklitsch (1939 and 1943), Meyer - Peter and Meyer - Peter and Muller (1948), Kalinske (1947), Blench (1966), Engelund and Hansen (1967) and Ackers and White (1973) were not able to describe the Bragg gauge bedload data, using mean hydraulic geometry characteristics, even when the equations were modified.

The capacity loads, at the four main hydrometric stations above Glenmore Dam, have been calculated from daily discharges, for the period 1967 to 1981, using the formulae which best described the measured capacity loads at each station. The mean annual load capacities, over a 15 year period, are: Little Elbow gauge 783 t; Falls gauge 2335 t, Bragg gauge 13453 t; and Sarcee gauge 1013 t. The average annual bedload is approximately one - quarter of the capacity load which suggests that bedload is supply - limited.

Considerable amounts of bedload are stored in sediment sinks along the river channel. Major sinks occur at the mountain front, where a substantial decrease in slope occurs. In addition, sediment is stored in sediment waves, which may be enhanced by log - jams, in the lower foothills and plains. Apart from these sink areas, the Elbow River channel is degrading. The locations of these sediment sources and sinks have important implications for bedload transport because the wave greatly influences the local rate of bedload transport. Further, the slow migration of these waves downstream suggests that the bedload transport will vary with time at a point in the river system.

The bed of the Elbow River is armoured along the thalweg and the bars are largely paved. Transport tends to occur over the bars, rather than over the armoured thalweg. The major sediment supply is from the channel bars, river banks and limited riparian areas. The bar and bank deposits

generally have similar grain size distributions to materials which form the bedload (particles larger than 1 mm). However, at low discharges, transport involves finer bed material, in the 1 to 16 mm size range. This material is derived from bank erosion and truncation of the finer textured distal bar ends.

The application of bedload formulae in a gravel bed river requires knowledge of three conditions: the appropriate particle size, a definition of the active width of bedload movement, and the definition of the relevant hydraulic characteristics for the active width of the representative river reach. At discharges of the same order as the mean annual flood, the bed material in transit in the Elbow River had a similar gradation to that of bulk bar deposits. Thus, bulk bar deposits adequetly represent the appropriate bed material size distribution. The active width of bedload movement appears to be limited to the paved portion of the channel, rather than the armoured portion. The appropriate width of bed movement can be defined based on cross - sectional variations in bed material size (D_{90}) and Shield's stability analysis. In the Elbow River the active width is approximately half the surface water width. Mean reach hydraulics appeared to adequately represent the hydraulics of the active width of bedload movement in the Elbow River. However, it is expected that the representative parameters will change with time at a given location because of the passage of slow moving sediment waves.

6. SOLUTE LOAD

A. INTRODUCTION

Stream solute loads may be derived from sub - surface and surface erosion, as well as from atmospheric and other sources (Singh, 1970). Therefore, solute load is not derived exclusively from denudation of the rocks and regolith of the drainage basin. However, because the additional sources of solutes are relatively small, particularly in areas such as the Eastern Slopes of the Rocky Mountains (Singh, 1970), the total dissolved solids content in the streamflow is taken to represent primary denudation of the drainage basin (McPherson, 1975; Zeman and Slaymaker, 1978).

Solute samples were collected in the Elbow River basin as part of the suspended sediment sampling program, using depth integrating samplers. The solutes were analytically defined as the material passing through a Reeve Angel 930 AH filter under vacuum. A subsample of the solution was drawn off by pipette, evaporated in an oven at 105°C and cooled in a dessicator, to establish the total dissolved solids (t.d.s.) concentration (Guy, 1969).

Random samples, taken from the Little Elbow gauge, Falls gauge , Bragg gauge, and Sarcee gauge, samples in 1978 and 1979, were analysed for t.d.s. concentration. In addition unpublished Water Survey of Canada t.d.s. concentration data, which coincides with their suspended sediment data in the 1973 to 1979 period, were also used to

determine concentration - discharge relationships for Bragg gauge.

The t.d.s. concentration - discharge relationships were used to estimate daily loads at Bragg gauge in 1978 and 1979. The use of the mean monthly discharge instead of mean daily discharges to calculate month loads produce almost identical results (± 2%). Thus, monthly loads were calculated from mean monthly discharges for eight years in the 1969 to 1979 period, with coincident suspended sediment data, for each of the full - time hydrometric stations.

To calculate solute loads from the Little Elbow River basin in the pre 1978 period, monthly discharge was estimated based on relationships developed previously (Chapter 3 and Appendix 2).

B. CONCENTRATION DISCHARGE RELATIONSHIPS

Unlike suspended sediment, the solute concentration at a station has an inverse relationship with discharge (Figures 6.1 to 6.3):

$$\text{t.d.s.} = 231\ Q^{-0.196} (r^2 = 0.38) \quad \text{(Little Elbow g.)} \quad ..(6.1)$$

$$\text{t.d.s.} = 270\ Q^{-0.180} (r^2 = 0.70) \quad \text{(Falls gauge)} \quad ...(6.2)$$

$$\text{t.d.s.} = 320\ Q^{-0.189} (r^2 = 0.78) \quad \text{(Bragg gauge)} \quad ...(6.3)$$

$$\text{t.d.s.} = 405\ Q^{-0.238} (r^2 = 0.70) \quad \text{(Sarcee gauge)} \quad ...(6.4)$$

where: t.d.s is total dissolved solids concentration (mg/l) and Q is discharge (m^3/s).

The power functions of each of the concentration - discharge models are similar, with the exception of the

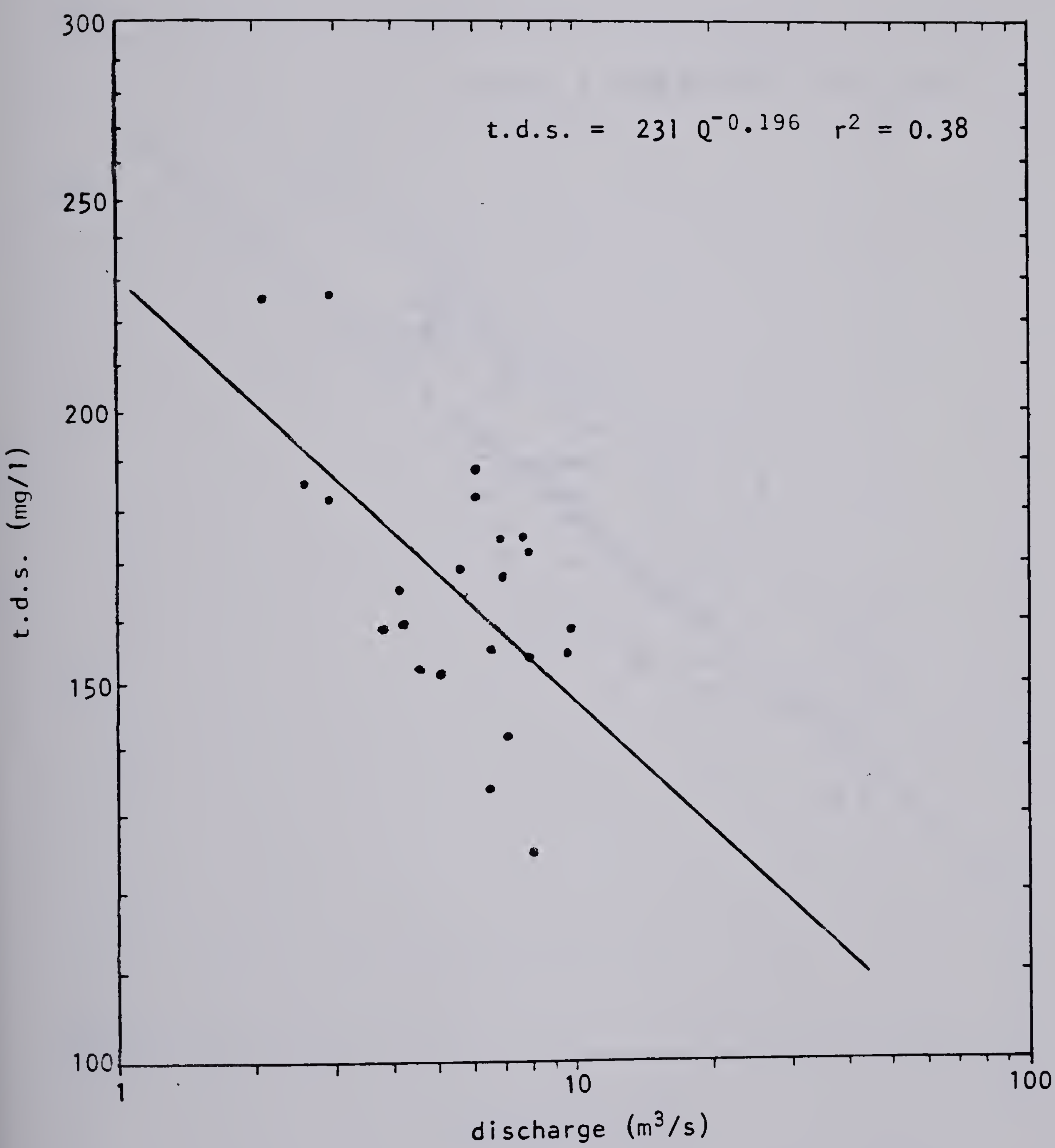

Figure 6.1 Little Elbow gauge total dissolved solids - discharge relationship, 1978 and 1979 data

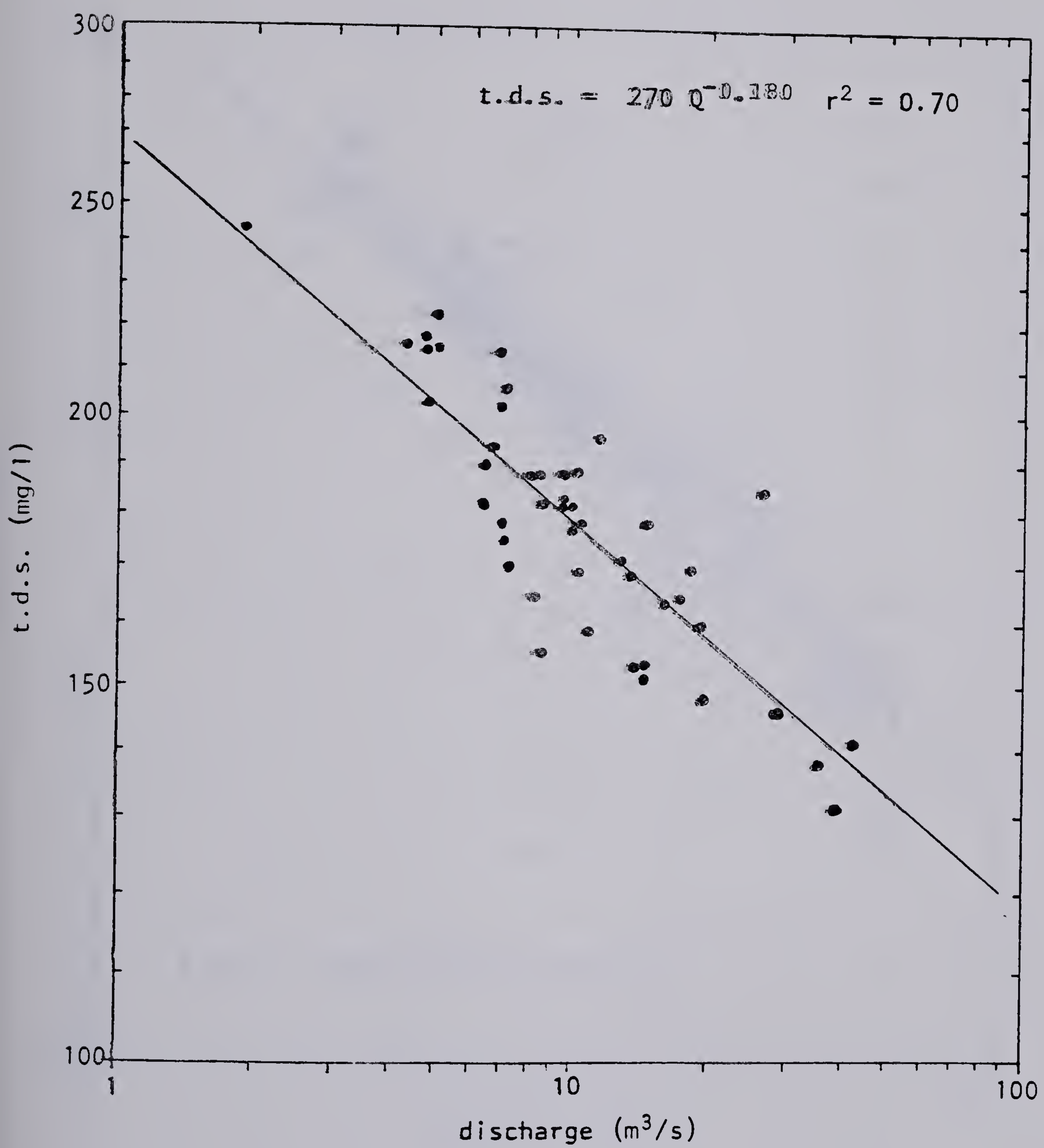

Figure 6.2 Falls gauge total dissolved solids - discharge relationship, 1978 and 1979 data

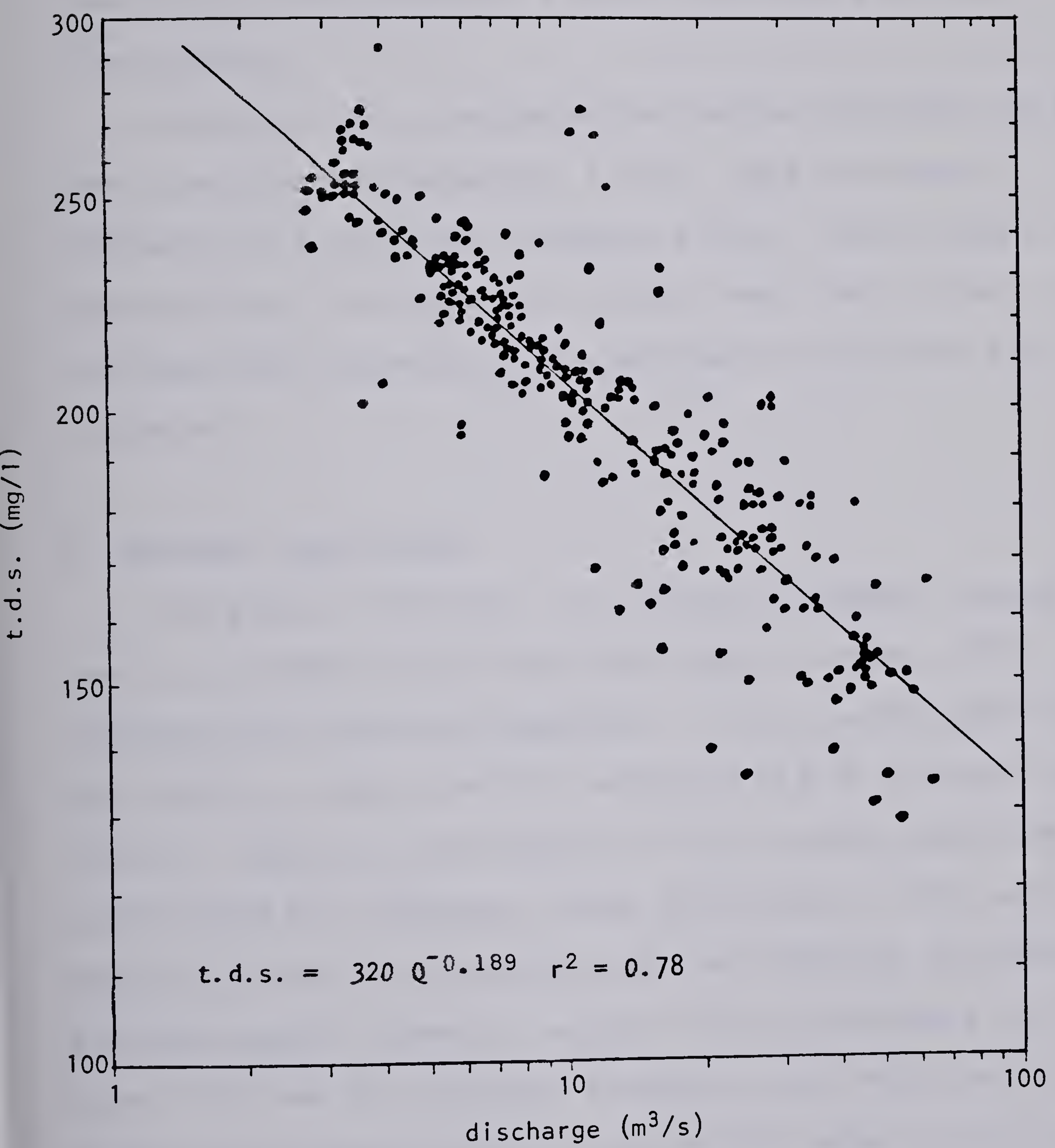

Figure 6.3 Bragg gauge total dissolved solids - discharge relationship, 1973 to 1979 data

Sarcee gauge model, which is steeper. The steeper Sarcee gauge rating curve is attributed to groundwater inflow in the lower basin below Bragg gauge (Appendix 3). A larger number of samples from the summer period, after groundwater flows stop, would probably flatten the solute rating relationship.

Although t.d.s. concentration varies inversely as a power function of discharge, a ten - fold increase in discharge, for instance, produces a one - third reduction in concentration. Consequently, solute load (the product of discharge and concentration) continues to increase with discharge.

C. TEMPORAL VARIATIONS

McPherson (1975:247), in a study of Eastern slopes basins, including the Elbow River basin, states: "The suspended and dissolved sediment - rating curves derived for the basins... were carefully analysed and no evidence was found to indicate that separate rating curves should be constructed for different times of the year." The large scatter in the relationships and the transient groundwater flow phenomenon, however, suggest this is probably not the case. This was not pursued, because a very detailed water chemistry approach, which is beyond the scope of this study, would be needed to explain temporal variations in concentration.

The average monthly and average annual loads for eight years of record, during the period 1969 to 1979, are shown in Table 6.1. The average annual load from this period is slightly less (2.5 to 5%) than for the 1935 to 1979 period load estimates (Table 6.1).

The average annual load at Sarcee gauge, in the period 1935 to 1979, has a coefficient of variation of 25%, which is a quarter of the variance of the suspended sediment load, and is less than the coefficient of variation of the discharge (33%).

Average monthly solute loads are relatively evenly distributed with time (Figure 6.4). The solute load distribution closely mirrors the runoff distribution. About 15% of the total annual solute load, and 13 % of the total annual discharge, occurs in winter. The greatest discharges and loads occur in May, June and July. On average 58% of the runoff and 52% of the annual solute load occur during this period (Figure 6.4).

The cumulative proportion of the total annual solute load transported with time increases at a declining rate. At Bragg gauge about 1.5% of the average annual load occurs in one day, and 10% occurs in about nine days (Figure 6.5).

D. SPATIAL VARIATIONS

Solute concentrations tend to increase downstream (Chapter 3 and Appendix 3). The increase in concentration downstream is paralleled by an increase in discharge

Table 6.1 Average monthly and average annual solute load estimates, Elbow River hydrometric stations

Station	J	F	M	A	M	J	J	A	S	O	N	D	Annual Tonnes
Little Elbow gauge[1]	488	460	430	513	1545	2936	1952	1200	1003	799	661	556	12543
Falls gauge[1]	1323	1248	1164	1392	4353	8242	5473	3316	2759	2161	1805	1512	34748
Bragg gauge[1]	1740	1732	2194	3309	8151	11893	7557	4685	4039	3284	2631	2097	53312
[2]	1745	1739	2127	3082	7927	11664	7967	5113	4601	3857	2662	2144	54628
Sarcee gauge[1]	2356	2450	3106	5159	9203	13441	8552	5131	4383	3599	3095	2488	62963
[2]	2452	2514	2914	4806	8845	13004	9053	6106	5387	4571	3556	2716	65925

[1]1969, 1971-1975, 1978 & 1979 data
[2]1935-1979 data

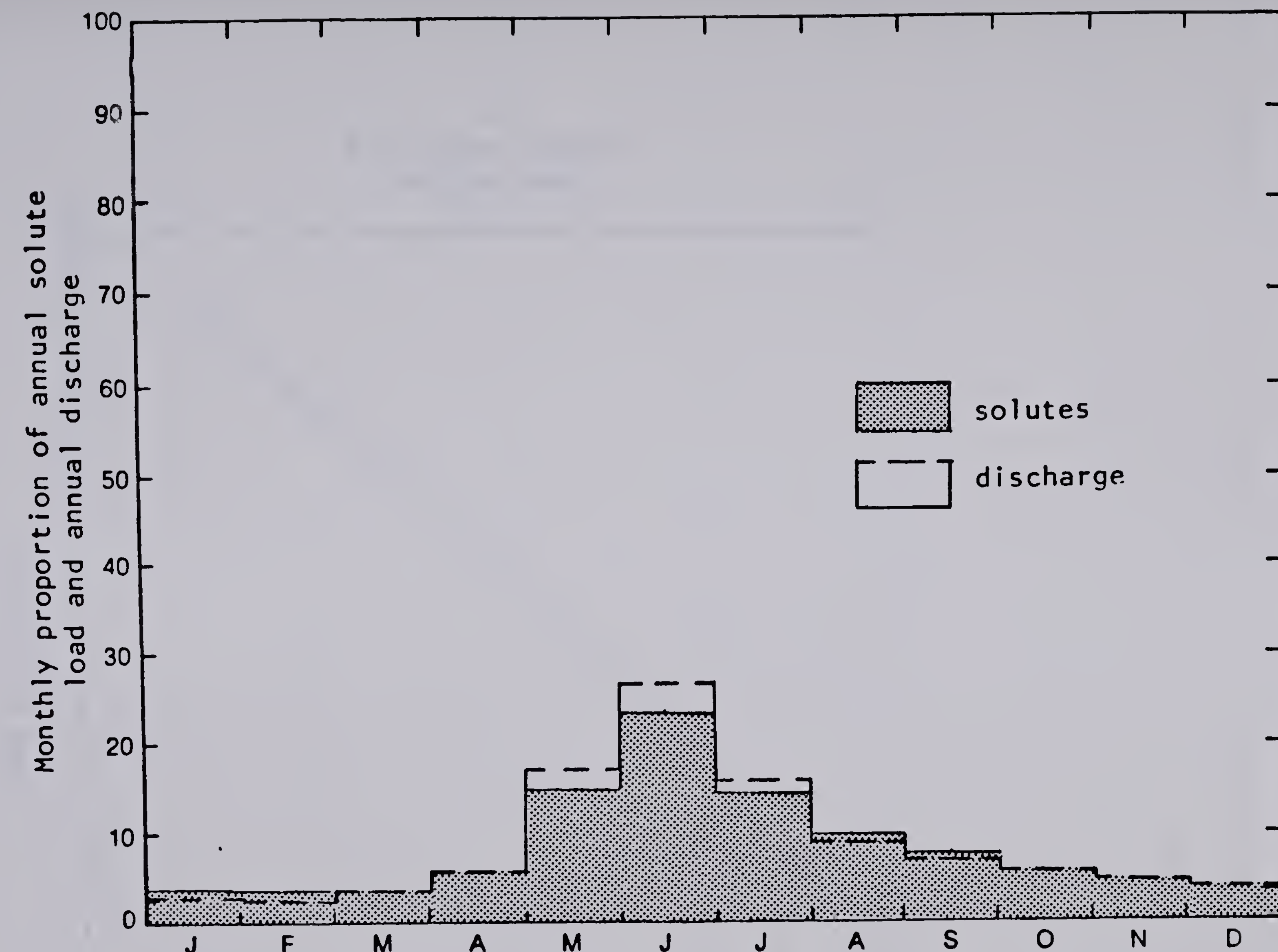

Figure 6.4 Monthly proportion of total annual solute load and annual discharge Elbow River at Bragg gauge, average of 1969 to 1979 data

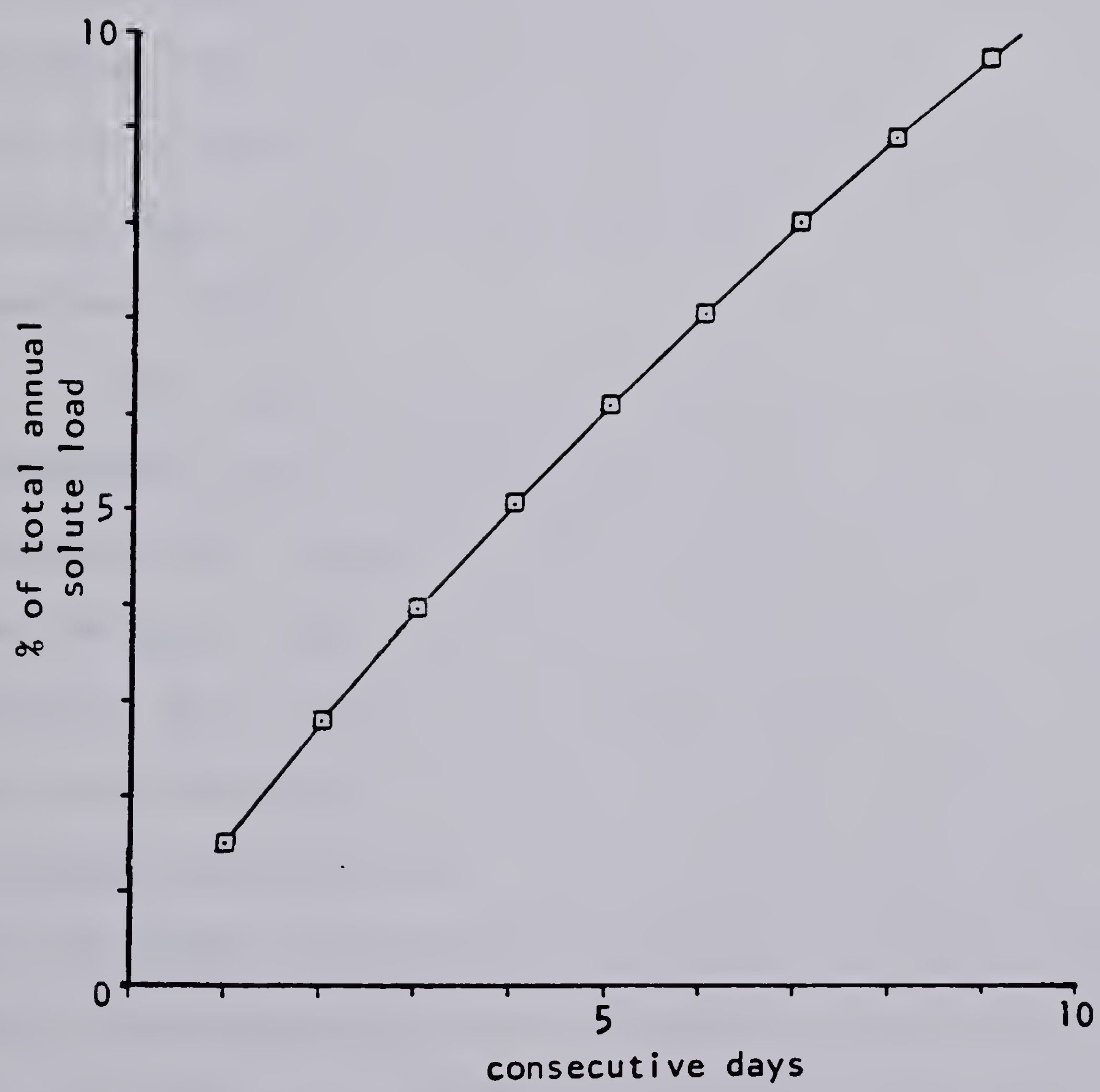

Figure 6.5 Cumulative percent of total annual solute load transported over time at Bragg gauge

downstream from the mountains to the foothills. The increase in discharge further downstream from Bragg gauge to Sarcee gauge is seasonal (Chapter 3 and Appendix 2).

Although solute loads increase downstream (Table 6.1), the sub - basin unit average annual loads decrease downstream, from 98 t/km^2 for the mountains, 65 t/km^2 for the upper foothills, 54 t/km^2 for the lower foothills, and 25 t/km^2 for the plains.

E. DISCUSSION

The laboratory procedures which were employed to determine total dissolved solids are meant to represent chemical weathering of the rocks and mineral soils of the drainage basin (Guy, 1969; McPherson, 1975; Zeman and Slaymaker, 1978).

In the surface water the predominant anion is bicarbonate and the predominant cations are calcium and magnesium (Environment Canada, 1980), which is to be expected given that the bedrock is primarily limestone and dolomite. As a result the average annual yield is greater than the continental average of 33 t/km^2(Livingstone, 1963 in Gregory and Walling, 1973). However, the loads are less than for other areas along the Eastern Slopes (McPherson, 1975). This may be due to a relative lack of karst system inflows in the upper Elbow River basin. This is a ground-water recharge area (Chapter 3 and Appendix 3).

The inverse relationship between solute concentration and discharge is attributed to dilution effects (Burt, 1979). During periods of low flow, runoff is generated from slow throughflow systems. These flows have a relatively high solute concentration, presumably because they achieve, or come close to achieving, solute saturation. The actual equilibrium concentration is relatively low in world terms, because of the low temperatures at these high elevations (Harmon et al., 1975). The short residence time of storm runoff further restricts dissolution, hence solute concentrations are relatively low from these inflows. In addition, overland flows from saturated near - channel margins, which add a significant component to flood generation, would have low solute concentrations, as do direct channel precipitation inputs.

The downstream variations in solute loads are to be expected in view of the hydrologic regime. Many of the mountain streams are ephemeral and their basins contribute sub - surface flows, via limestone and dolomite material, to the river network. In the foothills, which have a denser perennial stream network, the throughflow time is very rapid, which may not allow the throughflow water to achieve chemical saturation. Further, more direct precipitation and saturated surface runoff occur in this area. In the plains the load is derived primarily from regional groundwater inflows, with limited tributary runoff, and local ground-water input (Chapter 3 and Appendix 2). The near - surface

rock in the plains is sandstone.

F. CONCLUSIONS

Solute concentrations in the Elbow River basin are relatively low because of the generally low soil temperature of the mountains and foothills. Nevertheless, solute denudation, of the predominantly limestone and dolomite bedrock mountains and foothills, produces annual unit yields of over 98 t/km² from the mountains, 65 t/km² from the upper foothills, and 54 t/km² from the lower foothills. The 25 t/km² annual yield from the plains is derived primarily from regional groundwater inflow.

7. CONCLUSIONS

The two major controls of sediment transport in the Elbow River basin are exerted by the hydrologic regime and by the sediment supply regime. The importance of the former is derived from three facets of hydrology: (a) how runoff is generated; (b) where runoff is generated; and (c) the magnitude and frequency of discharge. Controls exerted by the sediment supply regime are divisible into two main components: (a) the characteristics of surficial materials; and (b) the availability of these deposits to fluvial processes. There are a number of related facets within each of these divisions.

The two main types of regolith which supply sediment to the river in the mountains are colluvium and lower valley wall glacial deposits. The colluvium is derived from weathering of limestone and dolomite. As a result little fine clastic material is produced. Runoff from ephemeral tributary streams provide most of the suspended sediment loads from the mountains in a short period in spring. The loads are limited by supply and the short duration of fluvial erosion. Surface erosion of finer textured glacial deposits on the lower valley walls is inhibited by coniferous forest vegetation protection and mantling of the toes of these deposits by bouldery lags. Infiltration and throughflow produce high solute loads at the expense of upland surface erosion. Surface erosion tends to be limited to areas where mass movements have occurred. The mass

movement scars intercept throughflow and permit surface erosion from snow melt and perhaps rainfall. Mass movements tend to occur in early spring, as the result of regolith saturation instability. The deposits may move downslope into the stream channel. These saturated deposits are rapidly removed by the first significant streamflows. As a result the bulk of the annual suspended sediment load is transported during a short period in spring. Bed load transport in the rivers of the mountains are less than the sustainable capacity load. This is mainly because a few tributary streams provide most of the sediment to the river channel. This material is stored in the river channel and removed during the limited high flow period.

The foothills of the Elbow River basin have extensive areas of highly erodible regolith. However, surface erosion is limited because the dense vegetation cover promotes sub - surface flow generation. Hence, solute loads are relatively high, but not as high as in the mountains, for two reasons. First, the regolith has less calcareous parent material. Secondly, the area is a groundwater recharge zone. The sub - surface flow systems that discharge into the stream network are usually local and very rapid, which inhibits solution. The major sources of suspended sediment yield are channel erosion and riparian erosion. Loads from these sources are supply - limited. Readily erodible toe accumulations are removed during the rising spring flood and a pavement develops which inhibits further erosion. As a result

sediment concentrations tend to be larger on the rising limb of the annual spring flood than following peak discharge. Bed load transport is supply - limited, thus, annual bedload transport is approximately one quarter of the annual mean capacity load. These loads can be predicted by only a few of the many available bedload formulae.

In addition to supply - constraints, fluctuations in the rate of bedload transport occur as the result of the mechanics of motion, and the passage of sediment waves. The former have a period of minutes, whereas the latter have a period of years.

The plains zone of the Elbow River basin also has potentially highly erodible surficial materials. However, the upland sediment yield component is minor for two reasons. Substantial runoff from the plains is rare and the transport network is inefficient because of internal drainage, low slopes and vegetated waterways. Also, the plains are frequently in a water deficit condition, hence there is little groundwater recharge and local discharge. The solute load increases downstream of the foothills is attributed to rapid, transient groundwater flows originating in the foothills and mountains which discharge in the lower basin river valley. The major sources of sediment are erosion of the river banks and valley walls. Toe protection tends to occur in these deposits as well. However, a large number of erosion sites do not have a protective pavement because the deposits are fine textured. Erosion from these

sites is sustained over a longer period. In addition, the valley walls may slump, as the result of rainfall, producing high suspended sediment concentrations without a significant increase in discharge. Bed load transport in the lower basin is supply - limited. As a result of bedload material supply limitations throughout the system, the Elbow River is tending to degrade.

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APPENDIX 1: CLIMATOLOGY OF THE ELBOW RIVER BASIN

INTRODUCTION

Although daily, seasonal, and annual fluctuations in air temperature, precipitation and other meteorologic elements are considerable, over a period of decades average conditions are recognised as representing the climate of a region. Major changes occur in the climate of the study area in both time and space. Temporal changes are addressed first.

Multiple glaciations of Pleistocene time attest to dramatic climatic changes in the recent geologic past. By Mid-Holocene time the local climate tended toward contemporary climatic conditions (Smith, 1979:48). More recently, somewhat less dramatic climatic changes resulted in neoglacial and warming cycles (Jackson, 1977). Even in historic time climatic changes are evident. Longley (1972:73-74) documents precipitation and temperature fluctuations over nearly a century. Although these fluctuations appear to be relatively small with respect to long term changes, significant changes in the hydrologic regime resulted (Appendix 2).

AIR MASSES

Three major air masses dominate the climate of the region. The variable extent and duration of these air masses determine precipitation and temperature from year to year. Air movement is predominantly westerly, but frequent shifts are associated with the passage of frontal cyclones.

In winter cold, dry polar continental (cP) air dominates the region. CP air is present 30 to 150 days per year in the Eastern Slopes area (Laycock, 1957:12) producing low temperatures and low to moderate precipitation (Tables 1, 2 and 3). The cP air mass does not produce precipitation _per se_ but causes cloud and light snow when maritime Pacific (mP) air overrides the cP mass (Laycock, 1957; Hare and Thomas, 1974). On 25 to 30 occasions per year on average, the Pacific air may descend as a warm, dry chinook which causes dramatic warming and removal, by sublimation, of snow from the mountain passes and prairies (Longley, 1972; Janz and Storr, 1977).

The rapid passage of westerly cyclones decrease in March or early April. This results in increased stability of the Arctic air and brings more settled cold weather. At the same time the main outflow of Arctic air shifts eastward and is replaced by warmer Pacific air streams which dominate the area by June (Hare and Thomas, 1974:84). Arctic outbreaks producing blizzards are rare. However, interactions of Arctic and Pacific air produce

Table 1 Mean Monthly Tempertures (^0C)

Station	Highwood Pass	Elbow Ranger St.			Moose Mt. L.O.		Glenmore Dam		
Record	1966-1973	1941-70	1978	1979	1978	1979	1941-70	1978	1979
Elevation	2202 m	1433 m			2434 m		1067 m		
Jan	-11.5	-11.9	-13.9	-14.6			-10.7	-13.5	-13.8
Feb	-9.2	-8.3	-7.8	-13.3			-7.1	-9.1	-16.8
Mar	-8.1	-5.9	-2.2	-1.3			-4.1	-1.7	-0.3
Apr	-4.7	0.8	1.9	-0.4			3.3	4.1	1.3
May	1.9	6.4	5.8	5.2			9.2	9.1	7.5
Jun	6.0	9.6	10.5	10.2	6.2	6.5	12.8	14.5	13.6
Jul	9.0	12.8	12.7	13.1	9.1	9.9	16.2	16.9	16.9
Aug	10.6	11.7	11.1	12.7	6.9	9.0	14.9	14.7	15.8
Sep	4.0	7.2	9.3	10.4	3.8	6.7	10.4	11.2	13.6
Oct	-1.4	2.9	5.0	4.4			5.6	7.0	6.6
Nov	-6.9	-4.3	-6.6	-3.0			-2.2	-5.0	-0.8
Dec	-12.9	-8.0	-10.4	-4.4			-6.9	-9.1	-4.8
Average	-1.9	1.1	1.3	1.2			3.5	3.3	3.2

N.B: Glenmore Dam missing data taken from Calgary International Airport.

Sources: Climate of Alberta: Report for 1978, Alberta Environment.
: Monthly Record of Meteorological Observations in Western Canada.
: Smith (1979) Highwood Pass data.

Table 2 Extremes of Temperature for each Month of the Year 1978-1979 (^{0}C)

Station	Elbow Ranger St.				Moose Mt. L.O.				Glenmore Dam			
Elevation	1433 m				2434 m				1067 m			
Year	1978		1979		1978		1979		1978		1979*	
Month	max.	min.	max.	min.	max.	min.	max.	min.	max.	min.	max.	min.
Jan	8.0	-34.0	9.0	-34.0					6.1	-28.9	5.6	-32.5
Feb	15.0	-33.0	8.0	-39.5					11.1	-32.2	6.5	-33.8
Mar	15.0	-33.0	17.0	-27.0					17.2	-25.0	17.7	-23.1
Apr	18.0	-17.0	17.5	-25.5					18.9	-6.7	19.6	-14.5
May	23.0	-7.0	25.5	-8.5	11.0	-7.0	15.0	-9.0	23.9	-1.7	26.8	-2.5
Jun	26.5	-3.5	27.0	-5.0	19.0	-6.0	21.0	-6.0	28.9	1.7	29.9	-0.3
Jul	29.0	0.0	31.0	-1.0	21.0	-4.0	22.0	-2.0	30.6	4.4	34.3	2.9
Aug	28.0	-2.5	27.0	0.0	21.0	-4.0	18.0	1.0	31.1	1.7	28.4	4.3
Sep	26.5	-5.5	27.5	-4.0	17.0	-8.0	19.0	-3.0	27.2	-1.1	30.6	-0.5
Oct	22.5	-13.0	25.5	-16.0	13.0	-4.0	13.0	-8.0	23.9	-4.8	27.9	-11.7
Nov	20.0	-34.0	19.0	-23.5					20.8	-26.5	19.0	-18.6
Dec	9.0	-37.5	14.0	-36.5					8.0	-30.5	12.7	-33.1

* 1979 Glenmore Dam data taken from Calgary International Airport

Sources: Alberta Environment Climate of Alberta Report for 1978

: Monthly record Meteorological Observations in Western Canada

heavy spring precipitation (Janz, 1976; Thompson, 1976). This precipitation is particularly important for flood generation (Appendix 2).

The dominant air masses in summer are the Maritime Pacific and warm humid Tropical Marine air masses. The tropical high pressure cells become unstable as they reach the eastern slopes. The upslope orographic lift results in protracted periods of low cloud and light snow or drizzle (Reinelt, 1970). Similar conditions result from westerly wind cyclogenesis, but precipitation may be major if, for example, cold moist unstable maritime air is aloft (Reinelt, 1968:20). Maritime Tropical air intrusions in late spring or summer may produce intense rainfall of long duration (Laycock, 1957:13).

Convective precipitation contributes considerably to summer precipitation both as rain and as hail. Powell (1977) notes that thunderstorms occur in preferred areas. The Calgary area has 20 or more days of thunderstorms per year, the foothills 13 to 17, and the mountain valleys less than 10. He reports, however, that there are probably more thunderstorms at higher elevations in the mountains. Hail storms are generated in the foothills and migrate eastwards in a swath 2 to 5 miles wide and usually less than 50 miles long. Stationary storms are rare and most are short-lived (71% less than 10 minutes duration and 6% more than 30 minutes duration), and infrequent (less than two per year for any township map unit) (Wojtiw, 1977). The hail season lasts from May to September but 50% of hailstorms occur in July (Powell, 1977).

In September the upper westerlies freshen and frontal precipitation results from moist air being uplifted over the cold Arctic air, which has moved southward (Rheumer 1953, Hare and Thomas, 1974). Easterly winds from Hudson Bay may produce orographic precipitation (Janz and Storr, 1977). Arctic outbreaks produce cold dry conditions and their dominance marks the beginning of winter.

In the Elbow River basin spatial changes in climate are related to elevation, topography and movement of weather systems. These changes are reflected in air temperature, precipitation and evapotranspiration.

TEMPERATURES

Three weather stations are currently operated within the boundaries of the study area. In addition miscellaneous climatological observations have been made in the basin and in adjacent areas (Table 2).

Values of mean monthly temperatures and extremes of temperatures (Table 1 and 2) demonstrate that the area is markedly continental with dramatic seasonal temperature

Table 3 Normals of Soil Temperature, University of Calgary

Depth (cm)		Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Annual
5	AM	-6.3	-5.0	-2.4	1.6	7.7	12.6	14.8	14.1	9.2	3.3	-1.5	-4.4	3.6
5	PM	-5.9	-4.3	-1.4	6.4	13.7	18.9	21.3	20.1	14.3	6.3	-0.6	-4.0	7.1
10	AM	-5.4	-4.4	-2.1	2.0	8.2	13.2	15.5	15.3	10.5	4.5	-0.4	-3.6	4.4
10	PM	-5.5	-4.4	-2.0	3.7	10.6	15.5	17.9	17.0	11.9	5.2	-0.3	-3.8	5.5
20	AM	-4.1	-3.6	-1.8	1.8	8.4	13.6	16.1	16.1	11.8	5.9	0.8	-2.2	5.2
20	PM	-4.1	-3.5	-1.6	2.1	8.7	13.8	16.2	15.9	11.8	6.0	0.9	-2.2	5.3
50	AM	-2.4	-2.3	-1.4	1.0	6.6	12.3	14.8	15.5	12.1	7.3	2.6	-0.9	5.4
100	AM	-0.1	-0.7	-0.7	0.3	4.6	10.3	13.0	14.3	12.1	8.5	4.4	1.3	5.6
150	AM	1.7	1.0	0.5	0.7	3.1	7.7	10.5	12.1	11.3	9.0	6.0	3.2	5.6
300	AM	5.4	4.6	3.8	3.4	3.6	5.2	6.8	8.7	9.6	9.4	8.4	6.6	6.3

N.B.: Elevation 1112 m, Location 51^0 05' 114^0 08'
Source: D.W. Phillips and D. Aston 1979
Soil temperature averages 1958-1978, Environment Canada Cl 13-79

variations. The extreme temperature data indicates, however, that considerable variations in temperature may take place in any month with the passage of chinooks and cold fronts.

The frost free period is over 100 days at Glenmore Reservoir, about 80 days at Bragg Creek (Longley, 1972) and 10 to 44 days at Highwood Pass (Smith, 1979).

A few short-period soil temperature studies have been conducted in the area. In the adjacent Kananaskis Valley the soils of forested areas freeze in October or November and remain frozen until April at lower elevations and May at higher elevations. Mean monthly temperatures in the upper few centimeters of these soils vary from about -6 °C in January to about 16 °C in July (in Lester, 1976). The corresponding temperatures at 5 cm depth in the plains area are -6.3 °C to 21.3 °C (Phillips and Aston, 1979). The soil in the plains area is frozen from November to March in the upper 10 cm. At a depth of 20 to 50 cm the period of frozen soil is from December to March. The soil at 100 cm depth is frozen during the first three months of the year. Soils below 1.0 m depth have an average mean monthly temperature above freezing (Phillips and Aston, 1979), (Table 3).

PRECIPITATION

Table 3 lists data from sites within the Elbow basin. Annual precipitation totals tend to increase from east to west from about 450 to 900 mm (Figure 1). Considerable variation in recorded precipitation amounts occur in the mountains (Longley, 1972).

In winter, which is taken to be October to April inclusive, the plains receive about 150 to 200 mm, the foothills about 200 to 300 and the mountains about 300 to over 500 mm of precipitation (Powell, 1977). Precipitation tends to be greatest in the mountains in winter whereas peak precipitation occurs in May and June in the foothills and plains (Powell, 1977). Summer precipitation totals are about twice winter precipitation in the plains, about 1.5 times as great in the foothills and generally less in the mountains (Powell, 1977; Table 3; Figure 1).

The annual recorded snowfall is not necessarily available for spring runoff. Buckler (1968) shows that it is possible that the snow pack may disappear due to sublimation with some melt, even at high elevations, during the winter or early spring. However, an entirely new snow pack may form with frequent and commonly heavy spring snowfall (Buckler, 1968:8).

Single-event rainfalls may approach 150 mm along the Front Ranges due to upslope flows, such as those associated with cold lows (Nemanishen, 1977). The heaviest one to thirty day rainfalls occur in the foothills and decrease both to the east and west

Table 4 Monthly and Annual Total Precipitation

Station	Elbow Range St.			Moose Mt. L.O.			Glenmore Dam*		
	av.	1978	1979	av.	1978	1979	av.	1978	1979
Jan	25.7	16.2	18.7				18.8	19.1	19.4
Feb	33.0	24.4	11.9				18.8	10.8	8.0
Mar	40.1	32.2	14.2				20.6	6.6	10.1
Apr	56.9	56.2	86.6				33.8	79.2	37.0
May	84.6	131.6	28.2		147+	22+	50.0	95.9	97.3
Jun	127.8	57.6	21.5		55.8	23.5	80.8	37.5	47.5
Jul	73.7	142.6	62.4		66.5	38.1	70.6	49.9	43.7
Aug	84.6	84.6	68.5		59.3	73.3	60.5	75.3	21.4
Sep	50.8	71.5	15.4		74.7	9.3	33.5	69.3	14.5
Oct	33.3	18.2	28.4		33+	20+	15.2	8.6	19.8
Nov	30.0	59.6	12.8				18.0	19.1	4.3
Dec	27.2	7.4	33.6				17.3	1.5	15.0
Totals (mm)	667.7	702.1	402.2				437.9	472.8	338.0
Snow (w.e.)	250.5	199.9					145.7		

*Glenmore Dam data supplemented by Calgary International Airport data where necessary.

+Based on partial record

Average 1941-1970 Conditions

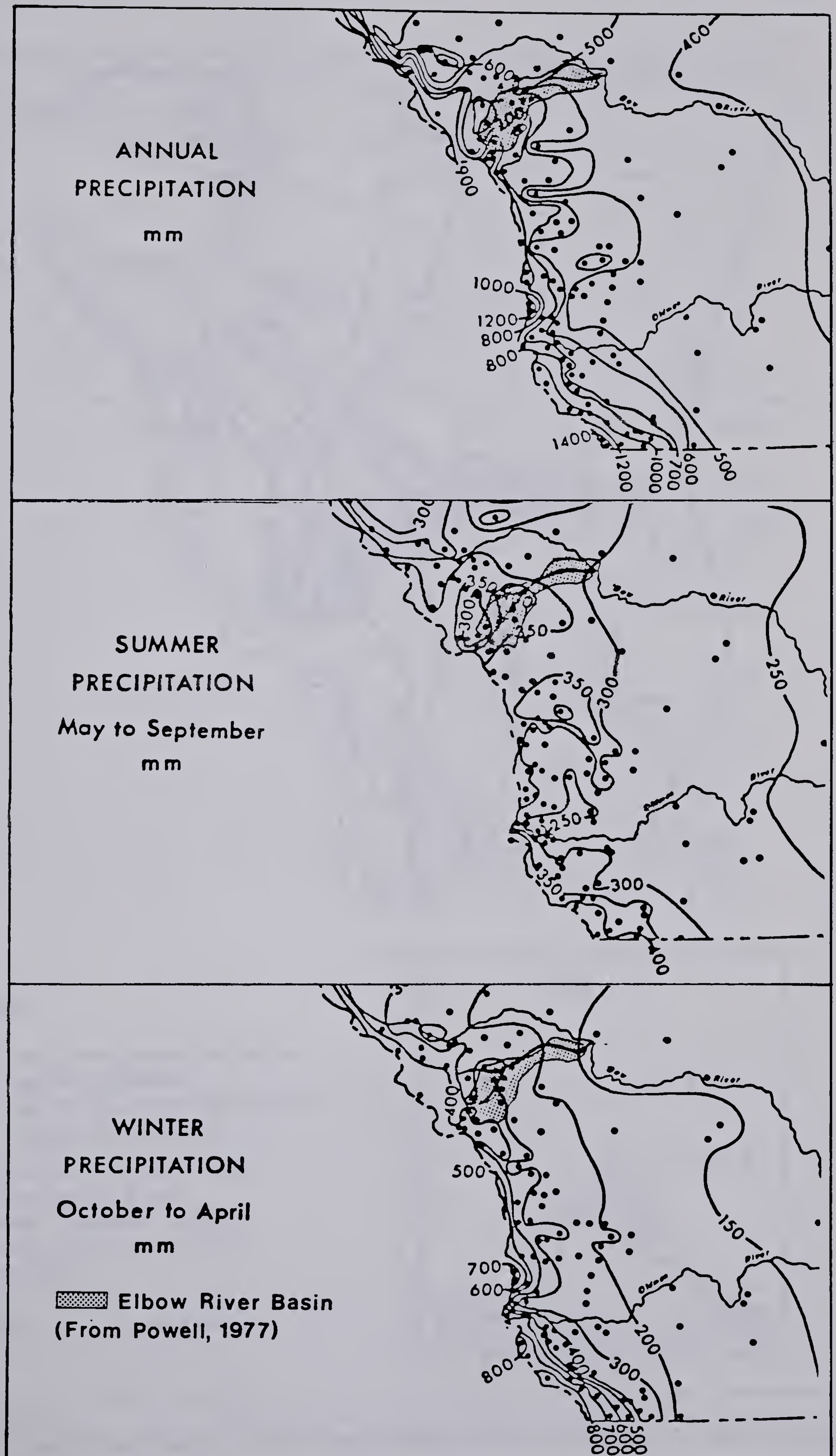

FIGURE 1 Elbow River basin area precipitation regime

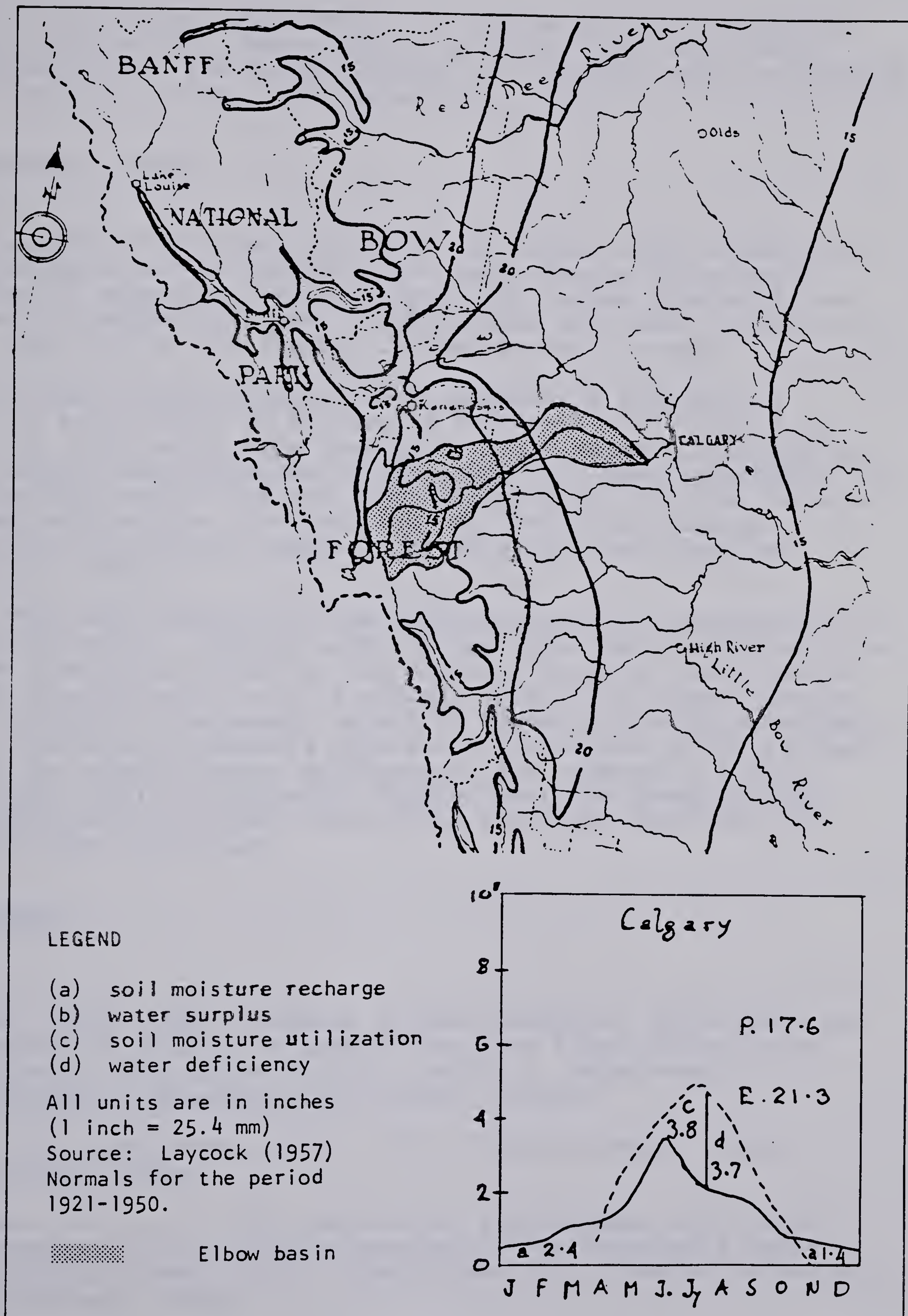

FIGURE 2 Elbow River basin area evapotranspiration and water balance normals

(Prairie Provinces Water Board, 1975; Powell, 1977). There is about a ten percent chance that a specific river basin with headwaters along the eastern Slopes will receive heavy stationary front rainfall in any year (Nemanishen, 1977).

ATMOSPHERIC LOSSES

Laycock (1957) has calculated the normal water balance for the mountains and foothills of the Saskatchewan River basin (Figure 2). Potential evapotranspiration ranges from well over 500 mm (20 inches) per year in the plains and lower foothills to well under 250 mm (10 inches) in some mountain areas.

"Actual evapotranspiration is apparently greatest in foothill areas where it is locally in excess of 20 inches per year. It is normally less than this in the warmer plains to the east because precipitation is deficient, at least seasonally. It is normally less in the more elevated areas to the west because of low temperatures, limited local soil moisture storage capacities, and local seasonal precipitation deficiencies..." (Laycock, 1957:4).

"The water balance of the Calgary station is representative of the foothills - plains margin areas which have relatively low winter precipitation and a June maximum. There is insufficient winter precipitation to fill storage capacities (other factors, such as runoff of snowmelt waters on frozen but unsaturated soil etc., should be considered separately) and there is no residual amount for runoff. Water deficiencies are moderately high." (Laycock, 1957:29; Figure 2). Further west the precipitation increases and the evapotranspiration decreases resulting in greater water surpluses.

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APPENDIX 2: FLOOD FREQUENCY ESTIMATES AND FLOOD GENERATION IN THE ELBOW RIVER BASIN

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INTRODUCTION

The Elbow River flows 120 km from the mountains, foothills and plains margin of the Eastern Slopes of the Rocky Mountains, into Glenmore Reservoir to meet the Bow River in Calgary (Figure 1). The reservoir has limited flood protection potential, at least as it is presently operated (Blench, 1965), and it is feared that a major flood would cause considerable damage, particularly within Calgary (Monenco, 1979). As a result, a number of schemes have been proposed to alleviate the flood hazard, such as increasing reservoir flood storage, increasing downstream channel capacity through the City of Calgary and building a flood control dam further upstream in the upper foothills (Blench, 1965; Monenco, 1979). An evaluation of the hazard, and of the abatement schemes, is contingent upon the elucidation of the flood generation processes and sources of runoff within the basin.

The hydrometric records for the Elbow River basin are rather a "mixed bag". Hence, it is necessary to evaluate the records at each of the stations in order to rationalize the data base for flood frequency analysis. Further, it appears as if the early 1900's were a period of considerably larger floods. Therefore, the records need to be extended, otherwise the more recent smaller magnitude floods will result in the underestimation of what a real, for example, 50 or 100 year flood would be.

This paper is therefore concerned with three problems - an evaluation of the data base, data extension of the short term stations and an evaluation of whether the runoff regime has changed with time.

HYDROMETRIC RECORDS - EVALUATION AND DATA EXTENSION

Data evaluation

The length and type of record of the Elbow River hydrometric stations is shown in Figure 1. The earliest station, the Elbow River at Calgary (station 5BJ1), renamed the Elbow River below Glenmore Dam in 1932, has operated since 1908. Following dam construction, the City of Calgary maintained a station above the dam (Elbow River above Glenmore Dam, 5BJ5) and provided Water Survey of Canada with daily discharge estimates for this station as well as for the station below the dam. The City estimated reservoir inflow by considering water use, an estimate of water release downstream, and the change in reservoir storage over a 24 hour period. Reservoir outflow was derived by simple subtraction. In 1969, Water Survey of Canada re-established the gauge, Elbow River below Glenmore Dam. In 1977, they stopped publishing the estimated inflows into Glenmore Reservoir because they questioned the accuracy of the daily discharge estimates (Spitzer, per comm.

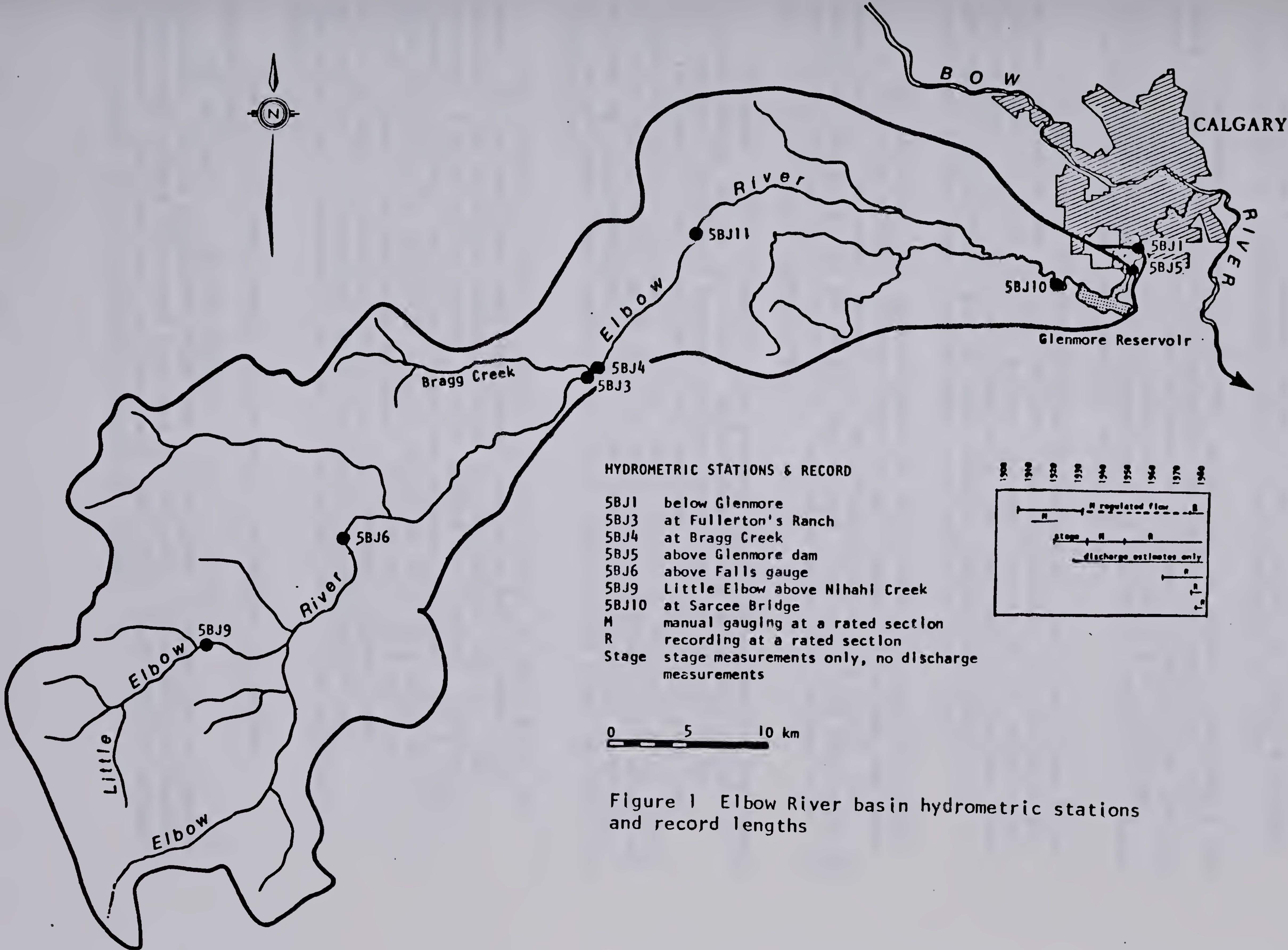

Figure 1 Elbow River basin hydrometric stations and record lengths

1981). Alberta Environment established a gauge at Sarcee Bridge (station 5BJ10) immediately above the reservoir in 1979.

The measurements at Sarcee gauge can be used to evaluate the most recent unpublished discharge estimates of the City. From Table 1 it is apparent that Glenmore inflow estimates are too low for the 1979-81 period. In fact, more than half the reported annual maximum mean daily discharges at Glenmore (5BJ3) are less than the discharge reported at Bragg gauge.

The accuracy of the discharge estimates for the Elbow River Above Glenmore Dam appear to be time dependent. In the 1951-63 period all but one of the annual floods increased in magnitude downstream. From 1935-50 about half the derived discharges show a decrease downstream, and in the 1964-81 period discharge is reported to decrease downstream two-thirds of the time.

The average loss downstream is 18 m^3/s and reported losses go up 56 m^3/s (1948). These downstream losses in annual maximum mean daily discharge appear to be unreasonable. In 1981, for example, measured Qd at Bragg gauge was 97.6 m^3/s and measured Qd at Sarcee gauge was 100.8 m^3/s. Instantaneous discharges were 124 and 121.5 m^3/sec respectively.

Dykema (1981) reports the average bankfull area in the reach around Bragg gauge is 145 m^2 and the average bankfull depth is 2.2 meters. Measured water surface profiles at 425 m^3/s by Hollingshead (1968) suggest the Manning 'n' is 0.025 and the slope is 0.0074 (Dykema, 1981). Mean velocity at bank full stage in this reach would be 5.79 m/s. This value agrees with measured values by Hollingshead (1968). The resulting bankfull discharge at the average cross section in this reach would thus be 840 m^3/s (29600 cfs).

The average bankfull depth in the Sarcee Bridge area is about 2.7 m and the average bank to bank width is about 60 meters. The average bankfull area is therefore 160 m^2. The channel slope of the lower basin, based on channel length and geodetic survey points at Twin and Sarcee bridges, is 0.0024. Mean velocity at bankfull stage would be 3.77 m/s with an 'n' of 0.025. Thus, the maximum bankfull discharge would be about 600 m^3/s. Between these two stations the channel shows some variabilty in character with reaches of aggradation and degradation. Overbank flow would be expected in the aggradation areas.

Over the three years of record, discharge increases downstream:

$$\text{Qd Sarcee} = 1.40 \text{ Qd Bragg}^{0.93} \quad (r^2 = 1.00) \qquad \ldots(1)$$

Further, it is apparent when the coincident discharges at Sarcee gauge and above Glenmore Dam are considered, that the estimated reservoir inflows are too low, and that this discrepancy appears to increase with increasing discharge (Table 1).

Table 1. A comparison of mean daily discharges of the Elbow Above Glenmore Dam (5BJ5) and at Sarcee Bridge (5BJ10), 1979-1981.

Year	5BJ5 Qd*	5BJ10 Qd	Ratio 10/5
1979	33.2	36.0	1.084
1980	45.7	52.9	1.158
1981	81.7	100.8	1.234
area	1210	1181	

*Data from the City of Calgary Water Works Department.

The Elbow River was gauged 1.45 km upstream of the present Bragg gauge (station 5BJ4, km 60.4) at Fullerton's Ranch (5BJ3) for the period 1914-23. During this period maximum annual mean daily discharge was reported on seven occasions (Table 2).

The Elbow River has been gauged at the town of Bragg Creek (station 5BJ4) since 1923. However, in the 1923-34 period discharge measurements were not made routinely, rather, stage was recorded during the high water period. On three occasions in 1923 discharge was measured at Bragg gauge. A stage discharge model, based on discharges of 11700, 1450 and 894 cfs, was used to predict discharges at Bragg gauge from the reported annual maximum and maximum mean daily stages. From 1935 until 1948 gauge heights were manually read once a day at a rated section and reported as mean daily discharge. As well, maximum instantaneous discharge was reported for four events which were the maximum instantaneous discharge up to that time. From 1949 until present stage has been recorded and discharge derived from the rating relationship at the station. On two occasions since 1950 (1966 and 1969) annual maximum instantaneous discharge was not reported (Table 3).

In the 1935-49 period four annual maximum instantaneous discharges were reported in the Annual Streamflow Summaries. In the "Historic Streamflow Summary to 1979" Water Survey of Canada present the annual publications maximum mean daily discharges, with the exception of the two highest mean daily discharges in the period where they have estimated a lower Qd from the reported Qi. Further, the 1947 Qd in the Historic Summary appears to be an error. The Qd reported in the Annual Streamflow Summary for 1946-47 is substituted (Table 1).

The Falls gauge hydrometric station (5BJ6, km 85) has continuously recorded stage on an annual basis since 1967 and the Little Elbow River gauge (station 5BJ9, km 106) is a seasonal gauge, which was established in 1978.

It is apparent that the most reliable hydrometric records in the Elbow River basin are recent. The most reliable long term station is the Elbow River at Bragg gauge (station 5BJ4). Therefore, Bragg gauge is used as the principal hydrometric station from which the data base is extended.

Data extension

To predict unknown annual maximum instantaneous discharges at Bragg gauge from mean daily discharges, regression models were developed. A log model provides the best description (r^2=.98) of the relationship between annual maximum instantaneous discharge (Qi) and corresponding mean daily discharge for the Elbow River at Bragg gauge (5BJ4):

$$Qi = 0.656\ Qd^{1.158} \qquad \ldots(2)$$

Table 2. Elbow River basin hydrometric records 1908-1934 and 1979-81 (m^3/s)

Year	Fullertons Ranch Qd	Fullertons Ranch Qi	Bragg gauge Qd	Bragg gauge Qi	Lower basin * Qd	Lower basin * Qi	Model A Qi	Model B Qi
1908			(109)	153 E	159	199 E	153	251
1909			(70.9)	92 E	94	119 E	92	132
1910			(18.9)	19.2 E	18.6	26.3 E	19.2	19
1911			(68.3)	88 E	89.5	113 E	87.5	125
1912			(88.1)	119 E	122	153 E	119	182
1913			(34.6)	39.2 E	38.8	51.1 E	39.2	45
1914			(27.2)	29.5 E	28.9	39.0 E	29.5	32
1915		320	(220)	352 E	239	381	278	411
1916	86.2		92.8 E	124.5 E	146	183 E	141	226
1917	65.3		69.6 E	89.3 E	147	184 E	142	228
1918	30.4		30.9	34.9 E	35.4	47.0 E	35.8	41
1919			(57.6)	71.7 E	72.5	92.5 E	71.7	97
1920	50.0		52.6 E	64.5 E	67.7	71.1	67.1	89
1921	34.8		35.7 E	41.2 E	37.4	49.4 E	37.8	43
1922	27.0		27.1	29.9 E	26.5	36.0 E	27.1	29
1923			(209)	331 E	331	402	312	610
1924			39.6 E	46.5 E	59.5	76.6 E	59.3	76
1925			49.3 E	59.9 E	66.5	85.2 E	66.0	87
1926			(67.7)	87 E	88.1	112 E	86.6	123
1927			44.2 E	52.8 E	83.3	106 E	82.1	115
1928			51.9 E	63.5 E	100	126 E	97.9	143
1929			(232)	376 E	382	433	359	725
1930			(33.1)	37.2	30.6	41.1 E	31.1	34
1931			(22.4)	23.5 E	22.9	31.6 E	23.5	24
1932			354	619 E	530 E	714	492	1152**
1933			46.7	56	42.6	55.8 E	42.9	51
1934			38.5	44.5				
1979			32.1	38.4	36.0	41.3	36.4	41
1980			51.7	69.3	52.9	59.7	52.9	66
1981			97.6	124	100.8	121.5	98	144

*Lower basin stations: 1908-1931 5BJ1; 1932-1934 above Glenmore (5BJ3); 1979-1981 Sarcee (5BJ10).
E estimates explained in the text
Bragg gauge (Qd) are estimates from Qi at Bragg gauge ($Qd=1.568\ Qi^{.843}$ ($r^2=.98$)
Model A: $Qib = 1.135$ Qd lower basin$^{.968}$ ($r^2=.93$)
Model B: $Qdb = .846$ Qd lower basin $^{1.046}$ ($r^2=.89$)
$Qib = .656\ Qdb^{1.158}$ ($r^2=.98$)

**The 1932 Model B estimate would be 565 m^3/s using the Elbow below Glenmore Dam Qd = 311 m^3/s.

Table 3. Elbow River at Bragg gauge (5BJ4) Qi and Qd 1935-1981.

Year	Qi		Qd	
1935	74.4	E	23.6	
1936	30.4	E	27.5	
1937	94.3	E	64.8	
1938	82.1	E	77.9	
1939	101.6	E	74.8	
1940	22.5	E	21.2	
1941	53.1	E	44.5	
1942	212.7	E	155	
1943	52.7	E	44.2	
1944	24.6	E	22.9	
1945	146.7	E	107	
1946	50.3	E	42.5	
1947	51.0	E	43.0	
1948	310	E	183	E
1949	18.1	E	17.6	
1950	58.0	JUN 15	44.2	
1951	110	AUG 30	82.7	
1952	71.4	JUN 12	49.3	
1953	181	JUN 13	118	
1954	43.9	AUG 25	39.4	
1955	47.6	MAY 19	32.3	
1956	35.4	MAY 21	30.9	
1957	30.0	JUN 8	28.9	
1958	40.2	JUL 13	37.9	
1959	45.9	JUN 27	39.9	
1960	29.4	JUN 3	28.9	
1961	55.5	MAY 27	51.0	
1962	28.9	JUN 16	26.1	
1963	199	JUN 29	113	
1964	97.4	JUN 8	89.5	
1965	153	JUN 18	125	
1966	34.4	E	30.6	
1967	430	MAY 31 *	254	
1968	50.7	JUN 10	43.9	
1969	196.9	E	138	
1970	125	JUN 14	94.0	
1971	138	JUN 6	101	
1972	60.9	JUN 1	52.7	
1973	49.3	JUN 7	43.3	
1974	78.4	JUN 17	66.0	
1975	53.5	JUN 20	49.3	
1976	42.5	JUN 6	36.0	
1977	17.1	AUG 13	15.8	
1978	49.8	JUN 6	44.5	
1979	38.4	MAY 27	32.1	
1980	69.3	JUN 4	51.7	
1981	124	MAY 26	97.6	

Qi annual maximum instantaneous discharge (m^3/S)
Qd corresponding mean daily discharge
E estimated values ($Qi=.656\ Qd^{1.158}$)

and $Qd = 1.568\ Qi^{0.843}$ ...(3)

The model is based on 30 data pairs from 1950-1981 at Bragg gauge (Table 3). If Qd occurred more than a day away from the date of Qi, because annual maximum daily and maximum instantaneous discharge were generated in separate flood events, the mean daily discharge at Bragg gauge within a day of Qi at Bragg gauge was substituted.

The indirect approach to extending the record of Bragg gauge (5BJ4) back to 1908 using the record of the Elbow River below Glenmore Dam gauge (5BJ1) is necessary because the Elbow River was dammed at km 11.8 in 1932. Hence the period of coincident record of the stations (1935 to present) is not comparable because of flow regulation. Thus two approaches were pursued to relate the discharge at Bragg gauge (5BJ4) to the longer record of the Elbow River at Calgary (station 5BJ1). The first approach is based on 7 years of data from Fullerton's Ranch gauge (station 5BJ3, from 1923 to 1923), 13 years of record from Bragg gauge (5BJ4, from 1923 to 1932 and 1979 to 1981), 3 years of record from Sarcee Bridge gauge (5BJ10, 1979 to 1981) and the coincident record from the Elbow River Below Glenmore Dam (5BJ1). The second approach, B, is based on 43 years of data from the Elbow River above Glenmore (5BJ5). However, the mean daily discharges reported for the Elbow River above Glenmore Dam (5BJ5) are derived, not measured.

Approach "A"

Fullerton's gauge is located upstream of Bragg Creek tributary, hence to estimate the discharge downstream on the Elbow River at Bragg gauge (5BJ4) from Fullerton's gauge data requires an estimate of the addition of runoff to the mean daily discharge from the 55 km² tributary. The Elbow River has been gauged above Elbow River Falls since 1967 (Falls gauge, station 5BJ6, km 85). To estimate the contribution of Bragg Creek tributary to the annual maximum mean daily discharge of the Elbow River at Bragg gauge (5BJ4), the mean daily runoff from the Bragg gauge sub-basin on the day of the peak discharge at Bragg gauge, was calculated for the 1967-1981 period (Table 4). The Qd of the Elbow River at Falls gauge (5BJ6) was subtracted from the coincident Qd at Bragg gauge (5BJ4) for the period of coincident record, 1967-81. The residual, which represents the Bragg gauge sub-basin contribution to Qd at Bragg gauge (5BJ4), can be weighted to estimate the discharge from Bragg Creek tributary. A simple area weighting (55.04 km²/358 km² sub-basin area = 0.154) was used because the tributary is roughly aligned with the Elbow River river valley, is approximately the same length (21 km vs 25 km Bragg to Falls gauge) and traverses similar physiographic areas.

The discharge at Fullerton's gauge site in the 1967-81 period can be directly derived by subtracting the estimated Bragg

Table 4 Bragg gauge sub-basin runoff, 1967-1981

Year	Bragg Qd 5BJ4	Falls Qd 5BJ6	sub-basin Qd	Bragg Creek tributary
1967	254	105	149	22.95
1968	43.9	27.3	16.4	2.53
1969	138	57.8	80.2	12.35
1970	94	52.1	41.9	6.45
1971	101	49.0	52.0	8.01
1972	52.7	38.8	13.9	2.14
1973	45	31.7	13.3	2.05
1974	66	52.7	13.3	2.05
1975	49.3	39.4	9.9	1.52
1976	36	24.8	11.2	1.72
1977	15.8	13.6	2.2	0.34
1978	44.5	35.7	8.8	1.36
1979	32.1	23.2	8.9	1.37
1980	51.7	38.0	13.7	2.11
1981	97.6	65.3	32.3	4.97

Bragg Creek tributary Qd estimated by area weighting ($55/358$ km^2=.154).

Creek tributary discharge from the daily discharge at Bragg gauge (5BJ4) (Table 4). The relationship between the estimated discharge at Fullerton's and from Bragg Creek tributary for the 1967-81 period can be used to estimate the discharge at Bragg gauge (5bJ4) from known mean daily discharge at Fullerton's gauge (5BJ3) in the 1917-23 period:

$$Qdsub = 0.11\ QdFull. -2.88\ (r^2=0.96) \qquad ...(4)$$

The Bragg gauge sub-basin model is:

$$Qdsub = .036\ Qd^{1.524}\ (r^2= 0.95) \qquad ...(5)$$

The maximum instantaneous discharge can then be estimated using the Bragg gauge Qd-Qi models.

On three occasions in 1923 discharge was measured at Bragg gauge. A rating curve, developed from measured discharges of 11700, 1450 and 894 cfs, was used to predict discharges at Bragg gauge from the annual naximum and maximum mean daily stages reported at Bragg Creek from 1923 until 1934 (Table 2).

For the period 1923 to 1934 there is often confusion between annual maximum stage and the annual maximum mean daily gauge height. Annual maximum stage sometimes corresponds to the equivalent of the maximum instantaneous discharge stage, whereas on other occasions the annual maximum stage and the annual maximum daily gauge height, (which is equivalent to the mean daily discharge stage), are given as the same value on the same day. Qi stage and Qd stage are unlikely to be the same value given the well defined relationship between Qi and Qd (eqn. 2). The reported annual maximum stage was presumed to represent the annual maximum mean daily stage unless the former was reported to be larger than the latter. This approach seems reasonable when the discharges downstream at Calgary are considered (Table 2). However, the 1930 daily stage, with an associated discharge of 37.2 m^3/s, probably represents Qi stage, given the mean daily discharge at Calgary of 30.6 m^3/s. The same applies to the 1933 and 1934 floods.

During the period 1915 to 1932 four of the largest floods reported occurred at the lower basin hydrometric station (5BJ01, Figure 1). The 1915 flood discharge at Fullerton's Ranch gauge of 320 m^3/s (11300 cfs) was estimated from high water marks and slope area calculations. The addition from Bragg Creek is estimated to be 32 m^3/s based on the previously described model (eqn. 5). The estimated instantaneous discharge at Bragg gauge is thus 352 m^3/s and the corresponding Qi at Calgary was 381 m^3/s (Table 2). The maximum instantaneous discharge at the Bragg gauge in 1923 was published in the annual water data report to be 331 m^3/s (11,700 cfs). This discharge was in fact estimated based on slope area calculations at Fullerton's Ranch, with an addition of 700 cfs from Bragg Creek tributary (Spitzer, pers. comm., 1982). It thus represents the peak instantaneous discharge. The 1929 discharge at Bragg gauge is estimated to be 376 m^3/s (13277 cfs) based on the maximum instantaneous discharge at Calgary of 433 m^3/s, using the following relationship developed from 1915, 1920 and 1923 data (Table 2):

$QiB = 0.86\ QiC + 4.7\ (r^2 = 0.99)$ ...(6)

In 1932 the Bragg gauge was washed away at an estimated discharge of 382 m^3/s. The corresponding maximum instantaneous discharge above Glenmore Dam was 714 m^3s (25200 cfs). Equation 6 predicts the discharge at Bragg gauge to be 619 m^3/s based on Qi at Calgary. This appears to be reasonable given that the Qi would be 641 m^3/s if the gauge was washed away at a stage corresponding to the mean daily discharge. The maximum instantaneous discharge of the Elbow River above Glenmore Dam in 1932 was reported to be 714 m^3/s (25200 cfs). The corresponding mean daily discharge is estimated to be 560 m^3/s based on reported Qi - Qd relations of the Elbow River below Glenmore previous to dam construction and the Elbow River at Sarcee gauge:

$Qd = 0.776\ Qi + 5.4,\ (r^2 = 0.95)$ ...(7)

Since 1979 the Elbow River has been gauged at Sarcee Bridge (station 5BJ10, km 19.5, 1181 km ^{2}), which is immediately upstream of Glenmore Reservoir (Figure 1). The data from Sarcee gauge can be used to represent the non regulated discharge of the Elbow River below Glenmore (station 5BJ1, km 9.7, 1220 km^2)(Table 2).

Nineteen years of record are therefore available to establish the relationship between annual maximum discharge at the Elbow River below Glenmore (5BJ1, Qdg) and the Elbow River at Bragg gauge (5BJ4, Qdb) (Table 2). Two models were developed which produce almost identical results when the predicted Qd is converted to Qi:

$Qdb = 0.62\ Qdg + 10.6\ (r^2 = 0.96)$ ...(8)

and $Qib = 1.09\ Qdg - 8.6\ (r^2 = 0.96)$ ...(9)

Conversely:

$Qdg = 1.55\ Qdb - 10.3\ (r^2 = 0.96)$ ...(10)

The twelve unknown discharges at Bragg gauge in the 1908 to 1935 period were estimated using the Qd Glenmore - Qi Bragg model (Table 2).

Approach "B"

Estimates of reservoir inflow made by the City of Calgary can be correlated with the measured discharges st Bragg gauge. The relationship can then be transferred 2.15 km downstream to utilize the pre-dam construction mean daily discharge record (1908-32) at 5BJ1, the Elbow River at Calgary. However, Qd result from different events on seven occasions during this period. In these cases coincident mean daily discharges at Glenmore were substituted (Table 2).

The resulting model was used to predict the mean daily discharge at Bragg gauge (5BJ4) from reported Qd at station 5BJ1 for the period 1908 to 1932:

$Qd\ Bragg = 0.846\ Qd\ Glen.^{1.046}\ (r^2 = 0.89)$ ...(11)

Conversely:

$$Qd\ Glen. = 1.774\ Qd\ Bragg^{0.849} \quad ...(12)$$

The predicted discharge at Bragg gauge (Qdb) was converted to Qi using the Qd-Qi model from Bragg gauge (eqn. 2). The results are presented in Table 2.

An evaluation of approaches

Four data periods are used to generate flood frequency estimates. The data from each period is evaluated. The 1950-1981 flood frequency data is based on 30 measured instantaneous peak discharges and two predictions of Qi derived from Qd at Bragg gauge. The relationship between Qi and Qd is strong ($r^2 = 0.98$), thus the data base may be regarded as being of high quality (Table 3).

In the period 1935 to 1949 stage was generally measured daily at Bragg gauge and a mean daily discharge was calculated. As well, the annual maximum instantaneous discharge was reported for three of the larger floods of the period and for the first year of record (Table 3). The reported Qi and Qd values do not strictly conform to the Bragg gauge Qi-Qd model which is based on 30 data pairs. Based on the relationship of four known values of Qi and Qd in the 1935-49 period, the published Qd values are less then expected for discharges greater than 50 m^3/s and greater than expected for smaller discharges. This implies that the correction applied to the instantaneous discharge of a single stage measurement per day is dubious. However, there is insufficient information to attempt a correction of the data base. Hence, the published Qd values, which are based on an undefined instantaneous discharge, have to be accepted at face value and used with caution. The known Qi values were used when available.

The discharge estimates at Bragg gauge in the 1908 to 1932 period are the least reliable. However, there is no direct way of evaluating the decisions made in calculating Qi at Bragg gauge. The estimates can be tested indirectly by comparison with adjacent basins for similar periods of record.

Model A is based on measured discharges at Fullerton's Ranch gauge which is 1.5 km upstream of Bragg gauge, measured stage at Bragg gauge with limited rating information and measured discharges in the more recent period. The Fullerton's Ranch mean daily discharge data are based on two stage measurements per day at a rated section, hence they may be a more accurate reflection of Qd than the single stage Bragg gauge records discussed earlier. An estimate of Qd from Bragg Creek tributary was added to each Qd at Fullerton's to estimate the Qd at Bragg gauge (Table 2). These additions are relatively small and they are verified by measured tributary inputs in 1978 and 1979 (Table 4). Hence, the area weighting appears to be valid.

The Bragg gauge record in the period 1923 to 1934 is somewhat questionable. The stage discharge - relationship is well defined by three high discharges (11700, 1450 and 894 cfs) but there are no corrections available to compensate for possible channel changes. Bedrock controls, which protrude through the alluvium, would limit the amount of possible scour. The banks at Fullertons gauge were not over flowed by the 1915 flood (Qi 11300 cfs). Thus, the discharge estimates are probably reasonable. The 1979-81 records are based on measured discharges at Sarcee bridge gauge and can be used to represent the discharge of the Elbow River Below Glenmore Dam (5BJ1) in the pre-dam condition.

According to Model A, when Qd at Bragg gauge exceeds about 20 m^3/s the 425 km^2 Elbow River basin below Bragg gauge contributes significantly to the peak annual discharge at Glenmore. The 1967 flood, which is the largest recent flood at Bragg gauge, can be used to evaluate equation 10. The May 31 Qd at Bragg gauge was 254 m^3/s. The model predicts a Glenmore discharge of 383 m^3/s, thus, the lower basin must contribute 130 m^3/s to the discharge which would require 30 mm of runoff from the lower basin from an 87mm storm with two main rain days.

Model B is based on the Elbow River Above Glenmore Dam - Bragg gauge relationship for the whole of the 1935 to 1977 data period. Because the Above Glenmore discharge estimates are frequently too low, the transfer of the model downstream to use the Elbow River Below Glenmore (5BJ1) data in the 1908-32 period results in over-estimates of the discharge at Bragg gauge. Thus, the Glenmore model (eqn. 12) predicts major discharge losses downstream. For example, with a discharge of 254 m^3/s at Bragg gauge (May 31, 1967 Qd) the discharge at Glenmore is predicted to be 195 m^3/s. The mean daily discharge reported by the City is 199 m^3/s. The 59 m^3/s loss is unlikely in view of limited routing attenuation and the expected input from 87 mm of rain in the lower basin at the time of the peak.

The instantaneous discharge of the Elbow River above Glenmore Dam on June 2 1932 was 25200 cfs (714 m^3/s). The estimated mean daily discharge, based on measured Qd-Qi relationships of the lower basin gauges (Table 2), is predicted to be 19770 cfs (560 m^3/s). The Glenmore Reservoir reduced the peak at the Elbow River below Glenmore (5BJ1) to 433 m^3/s (15300 cfs) with a mean daily discharge of 331 m^3/s (11700 cfs). Model B predicts the instantaneous discharge at Bragg gauge to be 610 m^3/s (21520 cfs) using the below Glenmore Qd and 1152 m^3/s (40670 cfs) using above Glenmore Qd (Table 2).

DISCUSSION

The two approaches suggest opposite relationships between Bragg gauge and the lower basin flood discharges. Approaches A predicts that as the discharge from the whole of the Elbow River basin increases the contributing area expands eastwards. Approaches B suggests that the lower basin contributes to small magnitude floods (discharges less than 45 m^3/s) but diminishes the relative magnitude of the flood as the size of the flood increases. Both models are probably true to some extent. There would be some attenuation of the peak instantaneous discharge downstream from routing. But because the channel is very steep and relatively large, such losses would probably be minor.

The second contention is whether the lower basin below Bragg gauge can generate substantial flood flows. The physiographic divisions of Figure 1 show that approximately 20% of the basin below Bragg gauge is foothills, with similar physiologic characteristics to the lower Bragg gauge sub basin. Presumably, the runoff response of this area would be similar to that of the adjacent foothills. The runoff response of the remainder of the lower basin can be described by analogy with adjacent basins. The adjacent Sheep River basin has a number of features which are common to the Elbow River basin. Hence, there should be hydrologic similarity between the Elbow and Sheep River basins. The lower portion of each basin, below the respective gauges, have an average drainage density of about 2.2, and slopes, surficial materials and land use are almost identical. The hydrometric records of the Sheep River at Turner Valley (5BL14, 554 km^2) and near Okotoks (5BL12, 1500 km^2) show that substantial downstream increases in discharge may occur (Table 5). Part of the input is derived from Threepoint Creek (505 km^2), which is largely a foothills stream. Thus, the coincident input from Threepoint Creek, and from the Sheep River above Turner Valley have been subtracted to derive the net runoff from the 441 km^2 lower basin. This separation shows that the runoff from the plains area of the Sheep River basin is substantial (Table 5).

Substantial increases in discharge downstream are also reported from the plains area of the Highwood River (Table 6). In this case the basin is considerably larger and the analogy is not as obvious. The point is still the same however - the eastern foothills margin and plains area may provide substantial runoff during the annual maximum flood.

The eastward expansion of the contributing drainage area during higher magnitude floods is due to the runoff generating processes. Most of the annual maximum discharges in Eastern Slope basins occur during the spring (May and June in the Elbow River area). Snowmelt runoff usually progresses up basin from the plains to the foothills and mountains over a period of days to weeks. For smaller magnitude floods the upper foothills - mountain snow melt is the almost exclusive source of runoff. The

Table 5. Sheep River basin daily discharges 1912-1916

Year	Sheep R. at Turner Valley 5BL14 554 km^2		Sheep R. near Okotoks 5BL12 1500 km^2		Threepoint Ck. nr. Millarville 5BL13 505 km^2		net runoff from the lower basin 441 km^2
1912	51.8	Jul 8	133	Jul 8	25.5	Jul 8	55.7
1913	29.4	Jun 28	44.7	Jun 28	13.9	Jun 28	1.4
1914	19.3	Jun 4	24.2	Jun 4	2.3	Jun 4	2.6
1915	133	Jun 26	606	Jun 26	60.0	Jun 26	413.0
1916	200	Jun 28	224	Jun 28	32.8	Jun 28	-8.8

Annual maximum mean daily discharge at Turner Vally and coincident mean daily discharge at the other stations are used.

Table 6 Highwood River annual maximum daily discharges 1908-1919

YEAR	Brown's Ranch 5BL8 1200 km^2		High River 5BL3 1980 km^2		Aldersyde 5BL9 2340 km^2	
1908			260	Jun 2	412	E
1909			125	Jun 3	171	E
1910			48.7	May 26	54.9	E
1911			92.0	Jun 3	118	E
1912			190	Jun 16	303	Jun 16
1913	54.4	Jun 28	62.9	May 31	76.7	Jun 28
1914	75.0	Jun 3	54.4	Jun 4	60.6	Jun 4
1915	214	Jun 26	227	Jun 26	328	Jun 26
1916	228	Jun 28			289	Jun 29
1917	136	Jun 9			204	Jun 3
1918	78.2	Jun 10			88.9	Jun 11
1919	71.6	May 23			72.2	May 28

snow melt runoff is usually enhanced, to varying degrees, by precipitation.

Major floods are derived from heavy rainfall near the time of the spring melt peak discharge (Buckler, 1968). Heavy rainfall is generated by stationary front storms which centre along the eastern edge of the foothills. Precipitation decreases both east and west of this area (Nemanishem, 1977). The lower basins of the Eastern Slope rivers are within the heavy rainfall zone. Soil moisture contents would be high because of the previous snow melt and limited evapotranspiration (Laycock, 1957). Thus runoff coefficients would be relatively high. Nemanishem (1977) states that there is about a ten percent chance that a specific river basin with headwaters along the Eastern Slopes will receive heavy frontal rain in any one year.

A further major contention of the data extension approaches concerns the magnitude of the floods in the 1908 to 1932 period. To evaluate the approaches, and to determine the significance of hydrologic persistence, five data periods were subject to frequency analysis (Table 7). Both approaches suggest that the 1908 to 1932 period had significantly larger floods that the more recent record. However, the difference in the estimated increase in magnitude is very large. Model A suggests that a 100 year flood, based on 1908 - 32 data, would be about 2.2 times larger than a 1950 - 81 data period 100 year flood. Model B predicts an increase of 4.6 to 5.6 times for the 100 year flood for the 1908 - 32 period over the 1950 - 81 period.

A comparison of 1908 - 1932 and 1950 - 1979 period 100 year floods from several other Eastern Slope basins in south western Alberta suggests that the early period of record did have significantly larger floods than the more recent period (Table 8). The average increase in the 100 year return period flood from the 1950 -79 to 1908 - 32 period is 1.5 times. If the interior mountain Bow River at Banff record is ignored, because the Front Ranges shelter the basin from heavy frontal rainstorms (Nemanishen, 1977), then the average increase is 1.60 times between the two periods. The ratio of the adjacent Highwood basin is 2.0 times.

FLOOD FREQUENCY ESTIMATES

The data extension analysis suggest that approach A may be the most appropriate method with which to extend the records at Bragg gauge for flood frequency analysis (Table 2). Known, or estimated, annual maximum instantaneous discharge data were analysed using two and three parameter log normal distributions by moments and maximum liklihood procedures (Kite, 1977). The "best" estimate was chosen with consideration of the root mean square error between measured and calculated discharges, the value of the skew coefficient, and the size of the third

Table 7. Flood frequency estimates, Elbow River at Bragg gauge (5BJ4), annual maximum instantaneous discharge (m^3/s).

TP	Data periods						
Years	1950-81	1935-81	1908-81 A	1908-81 B	1908-32 A	1908-32 B	1967-81
1.01	18	17	18	18	19	19	16
1.05	23	22	22	22	23	23	22
1.11	28	26	26	26	26	28	28
1.25	36	33	33	34	34	39	38
2.00	63	59	61	68	67	94	71
5.00	119	119	131	166	164	283	143
10.00	171	177	202	275	275	523	208
20.00	231	248	292	421	427	877	285
50.00	328	365	446	688	708	1580	407
100.00	415	474	593	957	995	2344	517
Skew	0.10	-0.07	0.01	-0.03	-0.05	-0.25	0.10
error	29.3	13.1	20.8	45.5	45.5	45.5	50.5
A	11.6	12.5	14.0	16.4	16.4	16.2	7.6
proc.	max. lik.	max. lik.	max. lik.	max. lik.	max. lik.	max. lik.	max. lik.

Table 8 Selected Eastern Slopes hydrometric stations 100 year floods for the 1908-33 and 1950-79 periods. Daily discharge, m^3/s.

Station		area km^2	100 year flood estimate 1908-1932	1950-1979	ratio
5BB01	Bow River at Banff	2210	411	363	1.13
5DB01	Clearwater River near Rocky	3210	1477	937	1.58
5BL09	Highwood River near Aldersyde	2340	921	460	2.00
5CC02	Red Deer River at Red Deer	11600	2063	1443	1.43
5AD03	Waterton River near Waterton	614	618	451	1.37

Log Normal 3 parameter distribution estimates.

parameter in relation to the measured flows.

Flood frequency estimates for the extended data period 1908 to 1981 are presented in Table 9. Rather than creating synthetic data for the upstream and downstream stations and subjecting this data to flood frequency analysis, the relationship between discharges at the various stations was used to predict the flood frequency estimates (Figure 2). This results in some physically unlikely discharges for low magnitude annual flow events - the discharge at Falls gauge is larger than further downstream at Bragg gauge. The shorter periods of record suggest the low return period discharges should be larger (Table 7).

CONCLUSIONS

Four conclusions may be drawn from the above analysis. First, the mean daily discharge estimates of the Elbow River above Glenmore Dam (station 5BJ3), which are estimated by the City of Calgary Water Works Department, tend to substantially under-estimate the annual maximum mean daily discharge. The error increases with the magnitude of the discharge. From this it follows that model B produces significant over-predictions of discharge at Bragg gauge when the above Glenmore - Bragg gauge relationship is transferred downstream to use the measured Elbow River at Calgary record (renamed the Elbow River below Glenmore Dam in 1932) from the 1908 to 1932 period. Model B is the approach used by Alberta Environment, Hydrology Branch, to predict flood flows at Bragg gauge in the 1908 to 1950 period. The error of the estimates of the mean daily discharges of the Elbow River Above Glenmore can not be corrected directly because their accuracy appears to be time dependent. Further, the Elbow River Below Glenmore Dam record in the period 1935 to 1970 was derived by the City of Calgary based on estimated inflow and estimated water use by the city. Thus, the records are not independent, so they can not be used as a check.

The second major point is that the lower basins of Eastern Slope rivers may add substantially to the discharge of large magnitude floods which are generated primarily from the foothills and mountains.

Thirdly, the early 1900's appear to be a period of significantly higher runoff than subsequent periods. This is attributed to hydrologic persistence rather than to any long term changes in the character of the basin, such as extensive burning which occurred in the pre 1900's during railroad construction.

Finally, approach A, based on Fullerton's gauge measurements in the 1915 to 1923 period, stage measurements at Bragg gauge from 1923 to 1934 and stage - discharge measurements at Bragg gauge since 1935, is adopted as the model to predict the unknown 1908 to 1932 discharges at Bragg gauge. The resulting 100 year

Table 9 Annual maximum instantaneous floods frequency estimates Elbow basin hydrometric stations.

TP Years	Bragg gauge 5BJ4		Falls gauge 5BJ6	Little Elbow 5BJ9	Sarcee gauge 5BJ10
1.01	17.9	(17.5)	17.2	10.0	24.5
1.05	22.1	(22.5)	20.6	11.1	30.0
1.11	25.9	(27.0)	22.8	11.8	34.9
1.25	33.1	(34.5)	27.1	13.0	44.1
2.00	61.4	(61.0)	41.1	16.4	79.6
5.00	131	(119)	68.1	21.8	164
10.00	202	(174)	90.9	25.6	248
20.00	292	(240)	116.8	29.5	353
50.00	446	(347)	156.1	34.8	529
100.00	593	(445)	189.8	38.8	694

Qi Bragg gauge based on 1908 to 1981 data described in the text.
Falls gauge based on 1967-81 flood frequency estimates compared with 1967-81 and 1908-81 Bragg gauge estimates
Little Elbow gauge estimates = 2.024 Qi Falls$^{.563}$ (r^2=.93)
Sarcee = 1.560 Qi Bragg$^{.955}$ (r^2=.93)

Figures in brackets refer to an average of the 1935-1981 and 1950-1981 flood frequency estimates at Bragg gauge.

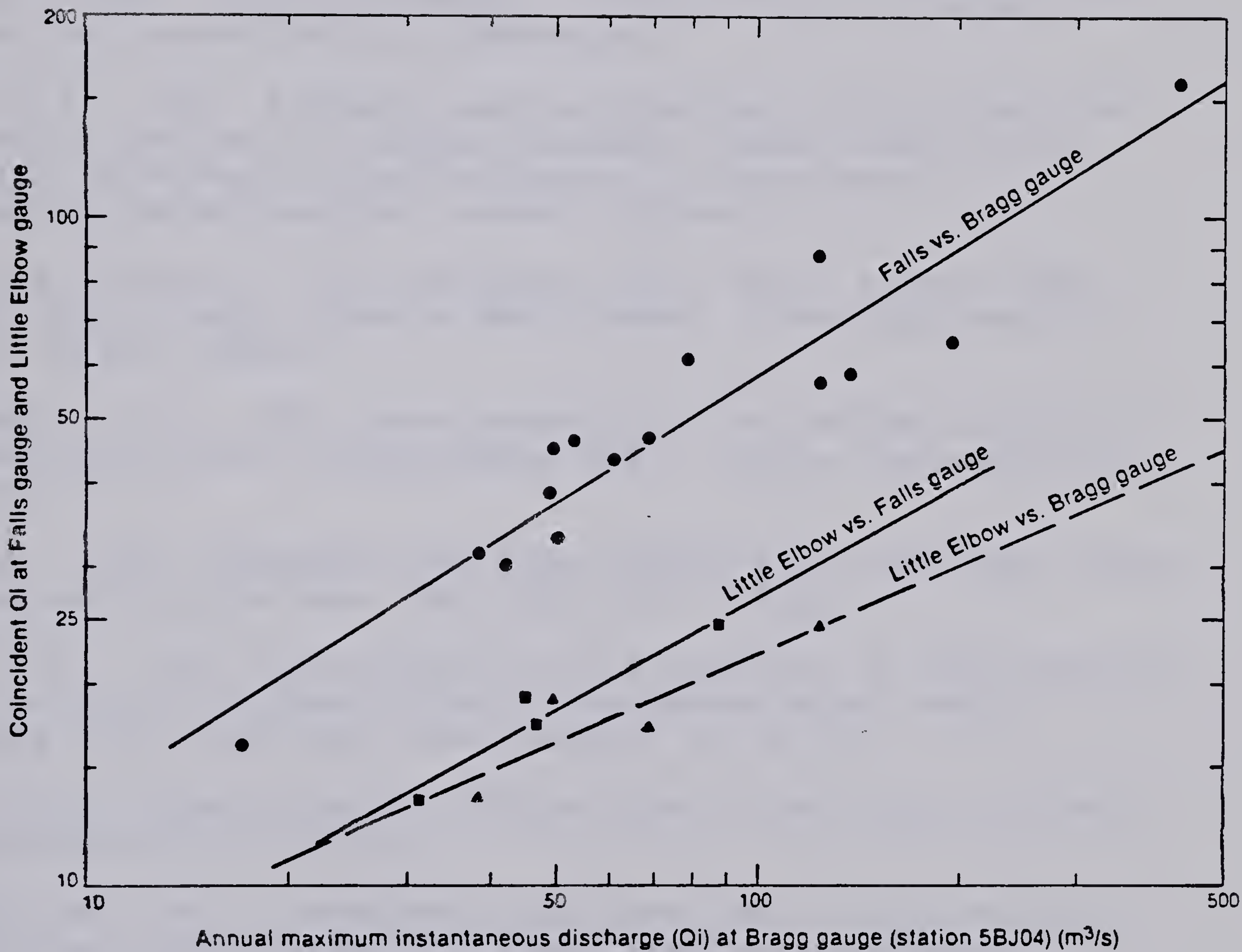

Figure 2 Elbow River basin hydrometric stations peak discharge relationships

return period flood at Bragg gauge is estimated to be 565 m^3/s (19950 cfs) using a log normal three paraameter distribution and maximum liklihood procedures. The results probably provide a conservative estimate of the 100 year flood because the 1908 to 1932 discharges appear to be somewhat overestimated when the ratio of 1908 - 32 to 1950 - 81 100 year flood at Bragg gauge is compared with adjacent Eastern Slope stations.

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APPENDIX 3: TRANSIENT GROUNDWATER FLOW PHENOMENON AT THE MARGIN OF THE EASTERN SLOPES OF THE ROCKY MOUNTAINS NEAR CALGARY, ALBERTA

H.R. Hudson

Civil Engineering Department Alberta Research Council Report No. U.62

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INTRODUCTION

Long term hydrometric records in the lower Elbow River basin suggest that discharge may seasonally increase or decrease downstream from the mid foothills at Bragg gauge (station 5BJ4, 793 km², km 60.4), to the Elbow River Above Glenmore Dam (station 5BJ5, 1210 km², km 19.5), which is in the plains (Figure 1; Table 1). The objective of this paper is to describe and explain the seasonal variations in discharge.

BASIN DESCRIPTION

The Elbow River flows about 120 km from the Front Ranges of the Rocky Mountains, through foothills and plains into the Bow River at Calgary (Figure 1). The general topographic form of NNW-SSE trending ridges and valleys, which decrease dramatically in elevation from the mountains in the west to the undulating plains in the east, is structurally determined (Evers and Thorpe, 1975). The form of individual ridges and valleys is attributed to subsequent glacial and fluvial erosion. The Elbow River and its upstream branches, the Upper Elbow to the South and the Little Elbow to the north, and some major tributaries (Bragg, Canyon and Prairie creeks) flow at almost right angles to the structual and lithologic trend of the basin (Figure 2).

The Elbow drainage basin is composed of three structural provinces: (1) the Front Ranges and (2) the Foothills of the eastern margin of the Cordillera and (3) the Plains of the Alberta syncline (Keating, 1975) (Figure 2 and 3).

The mountains and foothills are underlain by sedimentary strata originally deposited in marine or lagoonal environments on a Precambrian basement. These sediments were thrust northeast, relative to the passive basement, to produce a series of imbricated, northwest striking, generally southwest dipping, concave upward, locally folded and faulted thrust sheets (Figure 2 and 3; North and Henderson, 1954; Monger and Preto, 1972; Wheeler et al., 1972).

Older Paleozoic carbonate rocks were thrust eastward over younger Mesozoic clastic rocks in the front ranges (Wheeler et al., 1972). The thrust sheets are often steep to near vertical at their point of emergence (Douglas et al, 1970). The Paleozoic rocks tend to form peaks, cliffs and ridges while the recessive or less resistant clastics (shales and sandstones) of Mesozoic age are more frequent in the lower slopes and valleys (Milus et al., 1976).

The McConnell thrust fault divides the Rocky Mountain Front Ranges from the Rocky Mountain Foothills. In the western portion of the foothills thrust plates include Paleozoic strata, whereas

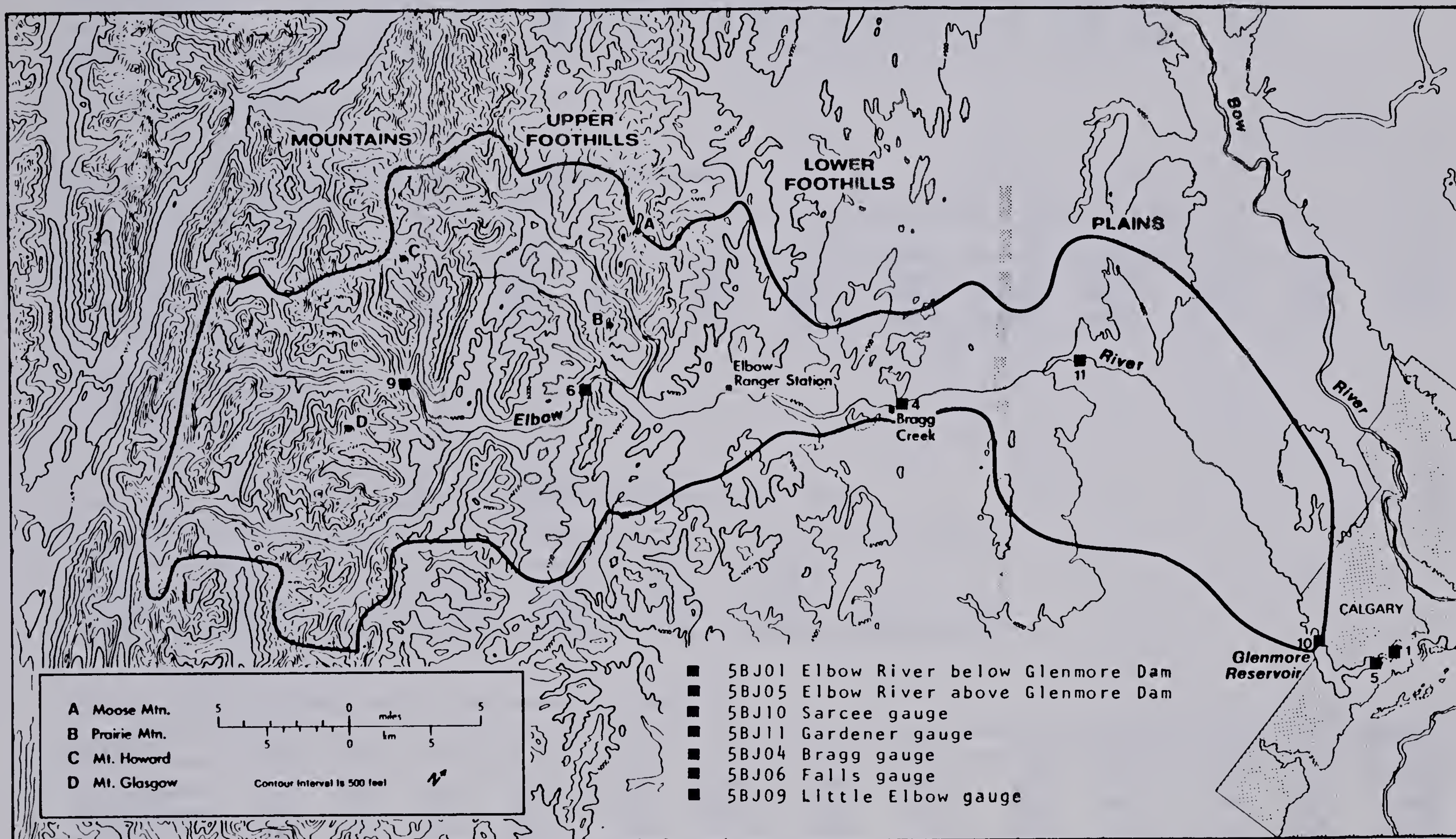

Figure 1. Elbow River basin topography and physiographic zones.

Table 1. Mean Daily Discharge, by Month, Elbow Hydrometric Stations, 1967 - 1979 (m^3/s)

Station-Month	J	F	M	A	M	J	J	A	S	O	N	D	ANNUAL
Elbow Falls	2.13	1.98	1.86	2.17	8.81	19.14	10.96	6.76	5.16	3.95	3.13	2.53	5.72
Bragg Creek	2.46[1]	2.45[1]	3.12	4.80	15.62	25.72	13.72	8.96	7.07	5.72	4.13[2]	3.17[2]	8.08
above Glenmore	2.76	2.97	3.71	7.03	16.16	28.29	14.42	8.92	6.50	5.50	3.90	3.02	8.60

[1] 3 years of data
[2] 4 years of data

Table 2. Mean Monthly Discharge, Elbow Hydrometric Stations 1967 - 1979 (da m^3)

STATION	J	F	M	A	M	J	J	A	S	O	N	D
Elbow Falls	5671[1]	4821[1]	4996	5625	23588	49575	29343	18110	13380	10595	8109[1]	6789[1]
Bragg Creek	6593[2]	5992[2]	8344	12438	41851	66652	36748	24001	18312	15322	10731[3]	8476[3]
above Glenmore	7387	7253	9946	18223	43277	73329	38624	23886	16851	14738	10102	8090

[1] 9 years of record
[2] 3 years of record
[3] 4 years of record

Table 3. Lower Elbow basin net monthly discharge, 1979 (da m^3)

Reach	A	M	J	J	A	S	O
Bragg to Gardner		7387	7550	3221	- 292		
Gardner to Sarcee		-3254	-2385	-2662	1497		
Bragg to Sarcee	3950	4133	5165	559	1205	- 208	- 188

Table 4. Mean net monthly runoff, Elbow sub-basins, 1967-1977

STATION	J	F	M	A	M	J	J	A	S	O	N	D	TOTAL
Falls gauge	5671	4821	5071	5648	24173	51766	30140	18767	13685	10738	8109	6787	185376
Bragg Creek	922	1171	3172	6764	18771	18421	7478	5867	4881	4803	2622	1687	76559
Glenmore	794	1261	1703	5811	333	3142	1006	-748	-1715	-803	-624	-386	14050
TOTAL	7387	7253	9946	18223	43277	73329	38624	23886	16851	14738	10102	8088	da m^3

LEGEND

Porcupine Hills Formation: *thick-bedded, cherty, calcareous sandstone; mudstone; nonmarine*
Tp

Paskapoo Formation: *thick-bedded, calcareous, cherty sandstone; siltstone and mudstone; minor conglomerate, limestone, coal and tuft beds; nonmarine*
TKp

Willow Creek Formation: *fine-grained, calcareous sandstone, bentonitic mudstone with white-weathering calcareous concretions; scattered thin limestone beds; nonmarine*
TKw

Brazeau Formation: *thick-bedded, chloritic and feldspathic sandstone and blocky grey mudstone; some tuff and thin coal beds; nonmarine*
TKb

Alberta Group: *fissile, silty shale; some thin-bedded, fine- to medium-grained, cherty sandstone (Blackstone Formation) – thick-bedded, well-sorted, quartzose sandstone; shale and carbonaceous shale; siltstone and thin coal beds (Cardium Formation) – dark grey fissile shale and siltstone; thin-bedded, fine-grained, glauconitic sandstone; thin beds of concretionary ironstone (Wapiabi Formation).*
Ka

Undifferentiated Mesozoic: *siltstone; dolomitic siltstone and limestone; silty dolomite, limestone, breccia and gypsum (Triassic) – fissile shale and siltstone; cherty and phosphatic dolomite and limestone; glauconitic shale and sandstone (Jurassic) – thick-bedded, fine- to coarse-grained, cherty sandstone interbedded with shale, siltstone and coal (Nikanassin and Kootenay Formations) – siliceous, calcareous sandstone; chloritic and feldspathic sandstone; carbonaceous and calcareous shale; shale and silty shale; some conglomerate; coal in central and northern Foothills; trachytic tuff, agglomerate in southern Foothills (Blairmore Group)*
Mz

Upper Paleozoic: *argillaceous limestone and dolomite, in part cherty and stromatoporoidal, in part coarsely biostromal, nodular and calcareous shale; aphanitic to finely crystalline limestone, dolomite-mottled limestone, and dolomite; bituminous shale (Upper Devonian) – fissile shale, siltstone, argillaceous limestone and cherty limestone; medium- to coarse-grained crinoidal limestone, cherty and dolomitic limestone; dolomite, cherty dolomite, anhydrite, shale and sandstone (Mississippian) – thin- and thick-bedded quartzose sandstone phosphatic quartzose siltstone, silty and cherty dolomite, chert and cherty carbonate (Pennsylvanian Permian)*
Pzu

Lower Paleozoic: *thick-bedded, quartzite and quartzose sandstone, with shale and limestone lentils (Gog Group) – thick-bedded dolomite and limestone; fine-grained, thick- to thin-bedded limestone and dolomite; calcareous and siliceous shales; local intraformational conglomerates (Middle and Upper Cambrian) – limestone, shaly limestone and shale, local intraformational conglomerate; medium-bedded, siliceous dolomite; clean white quartzite (Ordovician) – medium-bedded fine-grained dolomite (Silurian)*
Pzl

Rock unit boundary ....

Geology after R. Green (1970) Geological map of Alberta

Elbow River Basin

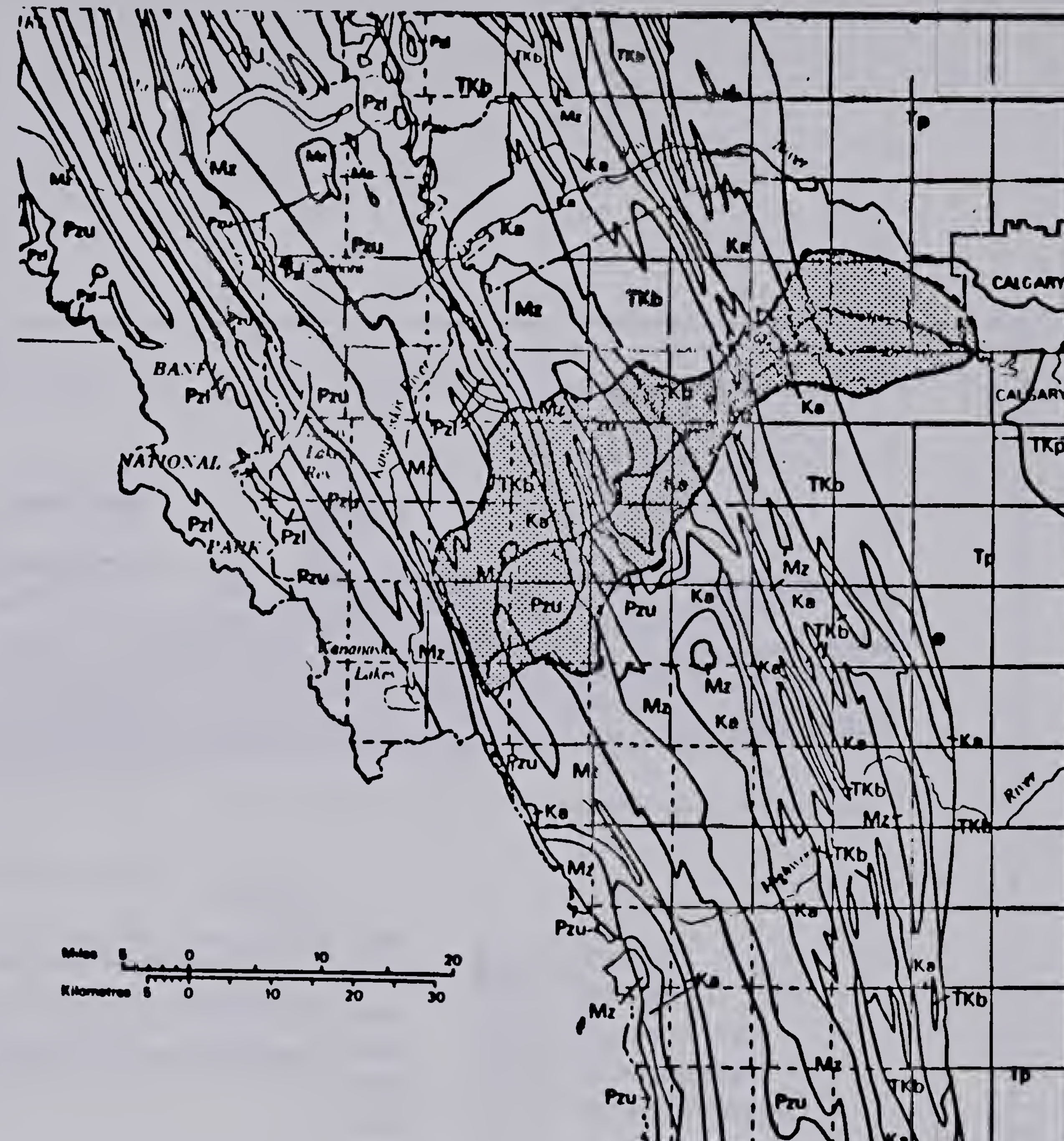

FIGURE 2 Bedrock geology of the Elbow River basin

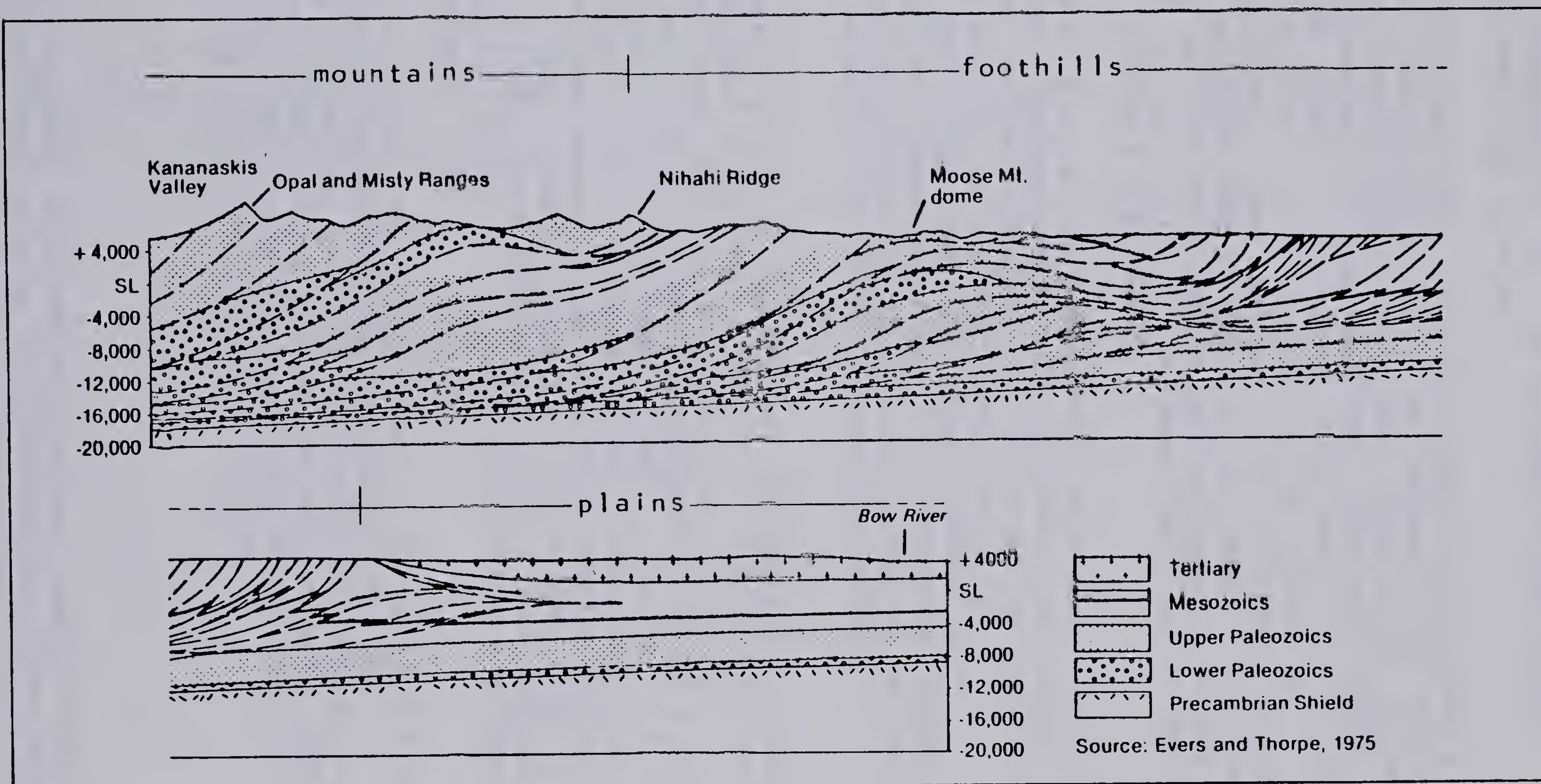

Figure 3 Elbow River basin structural cross - section

in the eastern portion of the foothills thrusting generally affected only the Mesozoic rock that consists of sandstones and shales which lie unconformably on Paleozoic Mississipian strata (Gordy et al., 1975; Keating, 1975). As a result the valleys further west are narrower and steeper.

The mountains, and the upper foothills west of the Falls gauge area (Figure 1), has structurally determined, high relief, ridge and valley topography. The valleys are broad and U shaped with numerous other glacial erosion features, such as cirques, horns, and aretes. The area is dominated by Paleozoic cliffs and an extensive colluvium cover. Till deposits are typically confined to valley bottoms, with the exception of cirque glacial deposits, and grade upslope into discontinuous veneers (Jackson, 1977). Glaciofluvial and fluvial plains and terraces line the valley bottoms. Most of the tributary streams have built fans which extend to the active flood plain.

Further eastward, in the mid and lower foothills, glacial erosion appeared to be limited mainly to truncation of spurs along the Elbow River valley (Seagel, 1971). Here the foothills are covered with extensive, relatively deep, morainic and colluvial deposits. The ridge tops are often covered with colluvial blankets which grade downslope into morainic blankets. The valleys are usually infilled with glaciolacustrine deposits or alluvium. These deposits are often covered by organic materials.

The eastern extent of the thrust sheet deformation is marked by the Alberta "syncline" (Keating, 1975:1). In the plains gently folded Tertiary sandstones and shales (Jackson, 1977) are overlain by extensive surficial deposits in the order of a few meters to tens of meters thick (Borneuf, 1980). The bedrock topography is very similar to the land surface topography except in the main river valleys where thick surficial sediments were deposited (Borneuf, 1980). The only bedrock units outcropping in the plains portion of the Elbow River basin are a few surface exposures and exposures along the Elbow River valley (Jackson, 1977; Ozaray and Barnes, 1978).

The two major surficial material types of the plains area are morainal blankets and glaciolacustrine deposits. The lacustrine deposits frequently overlie the morainic deposits. Alluvial deposits are confined to the valley bottom with local linear deposits across the plains to the Elbow River valley. Low discontinuous terraces flank the active floodplain.

BASIN HYDROLOGY

The long term, monthly, hydrometeorological records for the three long term unregulated gauging stations on the Elbow River illustrate the major features of the runoff regime (Figure 4).

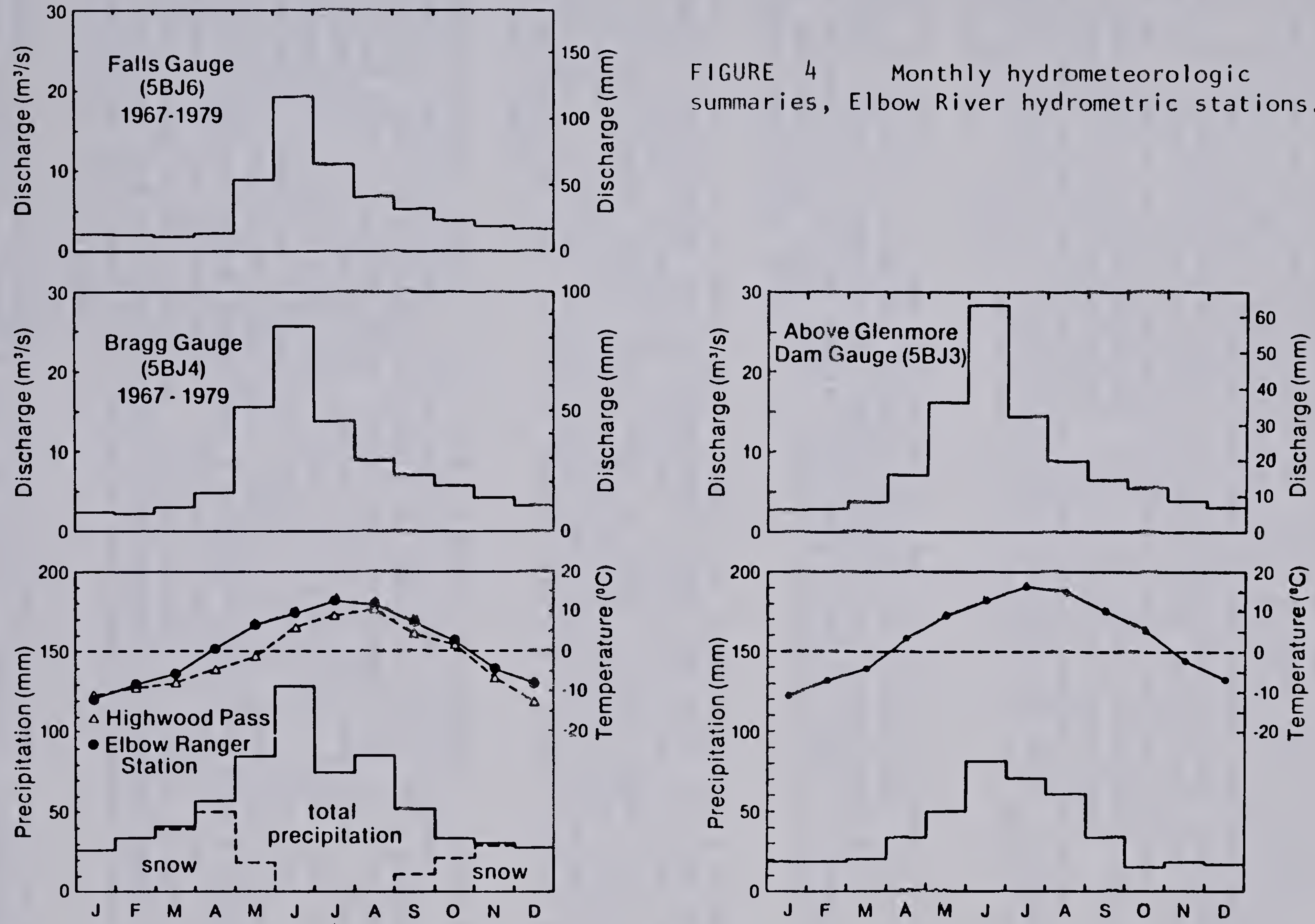

FIGURE 4 Monthly hydrometeorologic summaries, Elbow River hydrometric stations.

The three stations, Falls gauge (km 85, 435 km^2), Bragg gauge (km 60.4, 793 km^2), and the Elbow above Glenmore Dam (km 11.8, 1210 km^2), subdivide the Elbow River basin into three, almost equal, parts which represent the mountains and upper foothills, the mid foothills, and the plains, respectively (Figure 1).

During the winter months the river is ice covered and discharges are very low in relation to the remainder of the year (Table 1). About 13% of the total annual discharge occurs during the four winter months. Precipitation falls almost exclusively as snow and air temperatures are, on average, well below freezing (Figure 4). The storage of precipitation as snow, and the rapid release of the snow during melt, effectively redistributes several months of precipitation into the short snow melt period in the foothills and mountains. In the plains very warm temperatures, associated with chinooks, may remove the snow cover on several occasions in one year prior to spring melt. Chinooks may also occur in the foothills and mountains of the Elbow River basin (Buckler, 1968).

Spring runoff in the Elbow River basin is divisible into two distinct periods: an early spring period of relatively low discharges, where snow melt runoff occurs at progressively greater elevation up the basin over a period of days to weeks, and the main spring melt flood, which is dominated by runoff from the middle and upper basin.

When runoff from each sub - basin is separated by a simple input-output analysis, it is apparent that the bulk of the sub - basin total annual discharge occurs in April in the plains, and in May in the lower foothills, and in June in the upper foothills and mountains (Figure 5). Throughout these periods the bulk of the discharge is still derived from the foothills and mountains, but the proportion derived from these areas is less than at any other time (Table 1).

Approximately 60% of the total annual discharge of the Elbow River basin occurs in May, June and July. The spring peak discharge of the whole basin occurs in May or June. The monthly runoff from sub - basins show that, on average, 56% of the May runoff comes from above Falls gauge, 43% is derived from the Bragg gauge sub basin and less than 1% is derived from the lower basin between Bragg and Glenmore gauges (Table 2). The low proportion derived from the lower basin is, in part, attributed to the City Water Works Department underestimating the flood peak discharges for the Elbow above Glenmore Dam. In June, the flood generation zone shifts further up-basin and the Falls gauge sub basin produces 71% of the monthly discharge, the Bragg gauge sub basin produces 25%, and the Glenmore sub basin produces about 4% of the monthly discharge.

Summer recession usually begins in early June, and discharge decreases rapidly to reach a low in winter. Several, generally smaller, rainfall peaks can occur during the snow melt recession

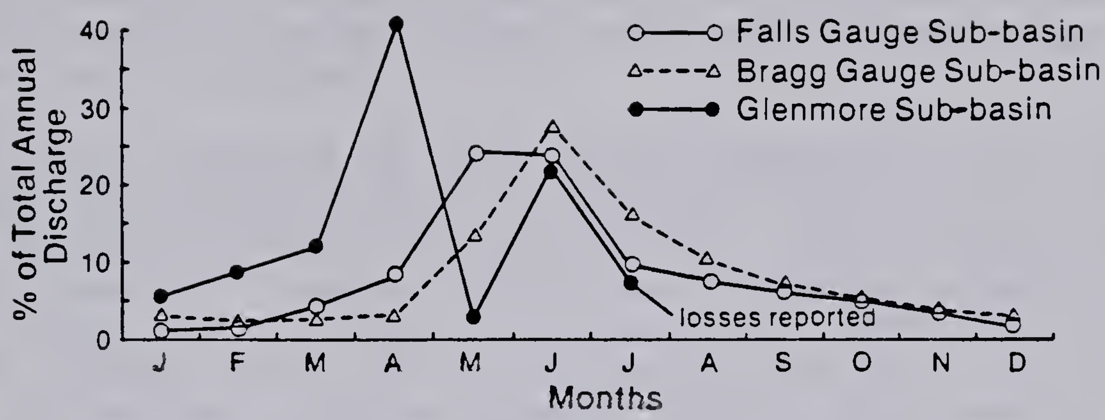

FIGURE 5 Mean monthly proportion of the total annual discharge from Elbow River sub-basins, 1967 to 1979

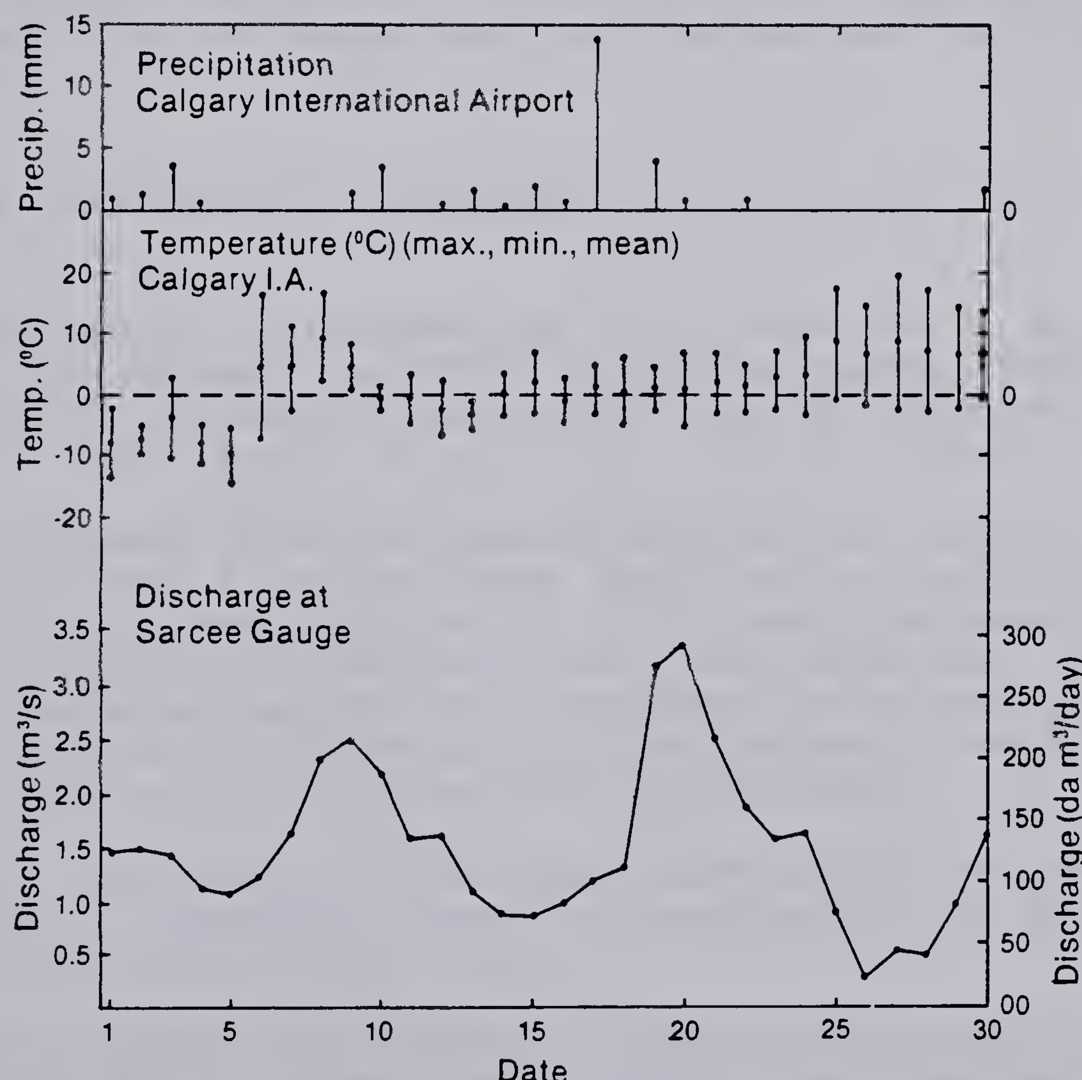

FIGURE 6 Net runoff, lower Elbow River basin, April 1979

period, which persists through July. During summer, and through winter, discharge increases downstream from the mountains and foothills. However, a major decrease in discharge occurs through the plains reach between Bragg gauge and Glenmore Dam during the summer and early winter.

Two points should be considered regarding this analysis. First, the discharge estimates made by the City of Calgary, based on water use, an estimate of reservoir outflow and the 24 hour change in reservoir water level, are of questionable accuracy (Spitzer 1981, pers. comm.), and have not been published by Water Survey of Canada since 1977. Secondly, the period of coincident records is limited. Winter runoff records are available for the lower basin for the period 1935 to 1977 (station 5BJ5). However, coincident winter records at Bragg gauge (5BJ4), are available for only January and February 1972, 1978 and 1979, and for November and December of 1971, 1977, 1978 and 1979.

To evaluate the apparent increases and decreases in discharge downstream, two hydrometric stations were established in the lower basin by Alberta Environment (Figure 1). Gardener Gauge operated from May 9 to September 3 and Sarcee Gauge from April to October inclusive, 1979. Gardener gauge was operated for the one year. Sarcee gauge has been maintained by Alberta Environment.

DOWNSTREAM INCREASES IN DISCHARGE

In the spring and summer of 1979 there was a net increase in discharge downstream from Bragg to Sarcee gauge (Table 3). The net increase in discharge is similar to the increase in discharge reported in other years (e.g.: 1970, 1972 & 1974).

The increase in discharge downstream is, however, not uniformly derived from the lower basin below Bragg gauge (Table 3). The net discharge figures in 1979 show that most of the discharge is derived from the river reach and sub - basin between Bragg and Gardener gauges and that there is a net loss in discharge between Gardener and Sarcee gauges in each month of record in 1979 with the exception of September.

The net increase in discharge generated from the Gardener sub - basin is thought to have two components (a) precipitation inputs and (b) groundwater inputs.

Laycock (1957: 33) suggests "The marginal and variable flow of the lower basin is received largely from snow melt waters and above normal spring rains in April and May." This situation appears to be the case over the long term (Table 4). and snow melt is thought to be the source of the excess runoff from the Gardener sub - basin in April 1979 because of the correspondence between discharge and meteorological conditions (Figure 6) and

because the groundwater input is estimated to be minimal during this period. This latter point is discussed in detail in a following section.

In the first seven months of 1979 approximately 200 mm of precipitation fell at the Sibbald Ranger Station, which is immediately north west of the Gardener sub - basin. For the three months of discharge records, May, June and July 1979, the excess runoff from the 78 km^2 sub - basin has been estimated, by separating the output from the sub - basin from the upstream input, at 232 mm of runoff. Thus, the precipitation input, per se, can not explain the estimated net runoff from the sub - basin, even when storage lags are considered and when substantial atmospheric losses (Laycock, 1957) are ignored.

Groundwater discharge and discharge from bank storage are thought to be the probable sources of the runoff from the Gardener sub - basin. Several lines of evidence support this hypothesis.

The graphical summary of mean daily flows at each of the Elbow hydrometric stations (Figure 7) shows that the discharge at Gardener gauge is consistently larger than Bragg Creek gauge in the summer of 1979. This increase in discharge downstream occurs throughout the diurnal discharge fluctuations. During May 1979, for example, the average increase in discharge downstream was 3.15 m^3/sec at peak instantaneous discharge and 3.35 m^3/sec at the daily minimum instantaneous discharge. Further, following each major rise in stage the difference between the two stations increases (Figure 7).

The consistently higher discharges reported at Gardener gauge is indicative of a fairly constant groundwater input, and the enhanced discharge following diurnal flow peaks and longer period rises in stage is what might be expected from recharge and subsequent discharge from bank storage.

The Gardener sub - basin overlies the structural edge of the foothills thrusts (Figure 1, 2 and 3) where numerous flowing wells, springs and seeps occur (Ozaray and Barnes, 1978; Borneuf, 1980). The discharge of springs and flowing wells in adjacent areas were found to be high (e.g.: near Harold Creek, which is 55 km n.w. a spring has a flow of 150 l/sec and there is a spring at the Bar K.C. club south of Bragg Creek town which has a flow of 42 l/sec) but is seasonally variable (Ozoray and Barnes, 1978; Borneuf, 1980).

Groundwater and bank storage components to the excess discharge of the Gardener sub - basin are suggested by two further lines of evidence - water chemistry and soil character.

Three groups of water samples taken at coincident parts of the respective hydrographs at Bragg, Gardener and Sarcee gauges in early spring, during the main spring flood and following the

Table 5. Water Quality, Lower Elbow Basin, May 1979.

Day Discharge	Element	Bragg		Gardner		Sarcee	
		ppm	load (g)	ppm	load (g)	ppm	load (g)
11-5	Na	2.199	17.02	3.085	34.74	2.982	30.51
	K	0.696	5.39	0.802	9.03	0.947	9.69
	Ca	49.90	386.23	50.98	574.03	43.51	445.11
	Mg	11.16	86.38	11.39	128.25	12.98	132.79
	Fe	0.021	0.16	0.052	0.59	0.915	0.15
Qi		7.74		11.26		10.23	
25-5	Na	2.934	67.01	3.538	96.55	1.94	49.72
	K	0.679	15.51	0.701	19.13	0.648	16.61
	Ca	45.11	1030.31	44.66	1218.77	50.71	1299.70
	Mg	9.727	222.16	9.525	262.67	10.99	281.67
	Fe	0.022	0.50	0.048	1.31	0.035	0.90
Qi		22.84		27.29		25.63	
28-5	Na	2.691	53.20	2.866	71.33	1.852	46.21
	K	0.597	11.80	0.661	16.45	0.646	16.12
	Ca	46.25	914.36	47.44	1180.78	44.88	1119.76
	Mg	10.15	200.67	10.34	257.36	11.53	287.67
	Fe	0.020	0.40	0.028	0.70	0.022	0.55
Qi		19.77		24.89		24.95	

N.B.: Qi is the instantaneous discharge at the time of sampling (m^3/s)

Table 6. Estimated Monthly Groundwater Discharge, Lower Elbow Subbasins, Spring and Summer, 1979

Subbasin	April	May	June	July	Aug.	Sep.	Units
Gardner		8455	8849	5278	1448		da m^3
		3.16	3.41	1.97	0.54		m^3/s
Sarcee	2624	5169	7880	3937	4044	1452	da m^3
	1.01	1.93	3.04	1.47	1.51	0.56	m^3/s

Table 7. Estimated Evaporative Losses in Monthly Discharge for Lower Elbow Basin Reaches, 1979 (da m^3)

Reach	M	J	J	A	S	O
Bragg to Gardner	27.4	29.3	46.3	35.5	29.4	16.9
Gardner to Sarcee	63.6	68.1	107.5	82.4	68.3	39.3
Bragg to Sarcee	88.8	95.0	150.0	115.0	95.4	54.8

Evaporation loss measured in a class A pan at Calgary International Airport. Loss estimated using a pan coefficient of 0.70 (following Schultz, 1973) for June to September values inclusive. May and October values estimated using Meyer's (1942) equation (r^2= .82 with measured values in adjacent months).

main spring flood were analysed. In addition published and unpublished data from Water Survey of Canada were reviewed.

The concentration or load of selected elements increases downstream from Bragg to Gardener gauge. The relative proportions of each of the elements generally do not change significantly between these stations (Table 5). Iron and sodium are exceptions.

The general maintenance of the relative proportion of elements may be expected because of the structure and repetition of lithologies upstream (Figures 2 & 3) and because of groundwater inflow upstream of Bragg gauge (Ozoray and Barnes, 1978). The augmentation of iron and sodium downstream are attributed to groundwater and bank storage discharge between Bragg and Gardener gauges.

Ozoray and Barnes (1978:25) used chemical composition changes to differentiate types of groundwater flow systems in the Calgary area. The surface waters of the Elbow River have a composition which suggests a local groundwater flow system where little SO^4 is available for solution. The groundwater emerging at Gardener gauge has an enhanced iron load. Since iron is relatively immobile (Stratham, 1977:61), the source of iron must be a slower, deeper groundwater system. A possible source of the iron in solution are iron-rich foothill springs, such as the Canyon Creek springs, which infiltrate into the stream bed a short distance downstream of their point of emergence in the foothills (Borneuf, 1980).

The point of emergence of the iron-rich groundwater appears to be close to the Gardener gauge because iron concentrations are higher there than further downstream where complete mixing may be expected to occur (Table 5: WSC, unpubl. data).

A bank storage component to the excess of discharge recorded at Gardener gauge is further suggested by the enhanced sodium concentrations in the river. MacMillan (1980) has described the soils of the Elbow River valley in the Gardener sub - basin area as sodium rich. Since sodium is mobile in the soil profile (Stratham, 1977:61), it might be expected to feature in the discharge from sodium-rich bank storage.

Increasing discharge downstream is thus apparently explicable in terms of discharge from bank storage and a fast, shallow, groundwater system in addition to precipitation in the Gardener sub - basin. However, the groundwater contribution appears to be highly seasonal (Table 6).

The daily discharge reported at Sarcee gauge is usually lower than the coincident discharge reported at Gardener gauge. However, when atmospheric losses are taken into account (Tables 7 & 8) the estimated gross discharge [1]

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[1]The estimated gross discharge from a sub - basin is equal to the

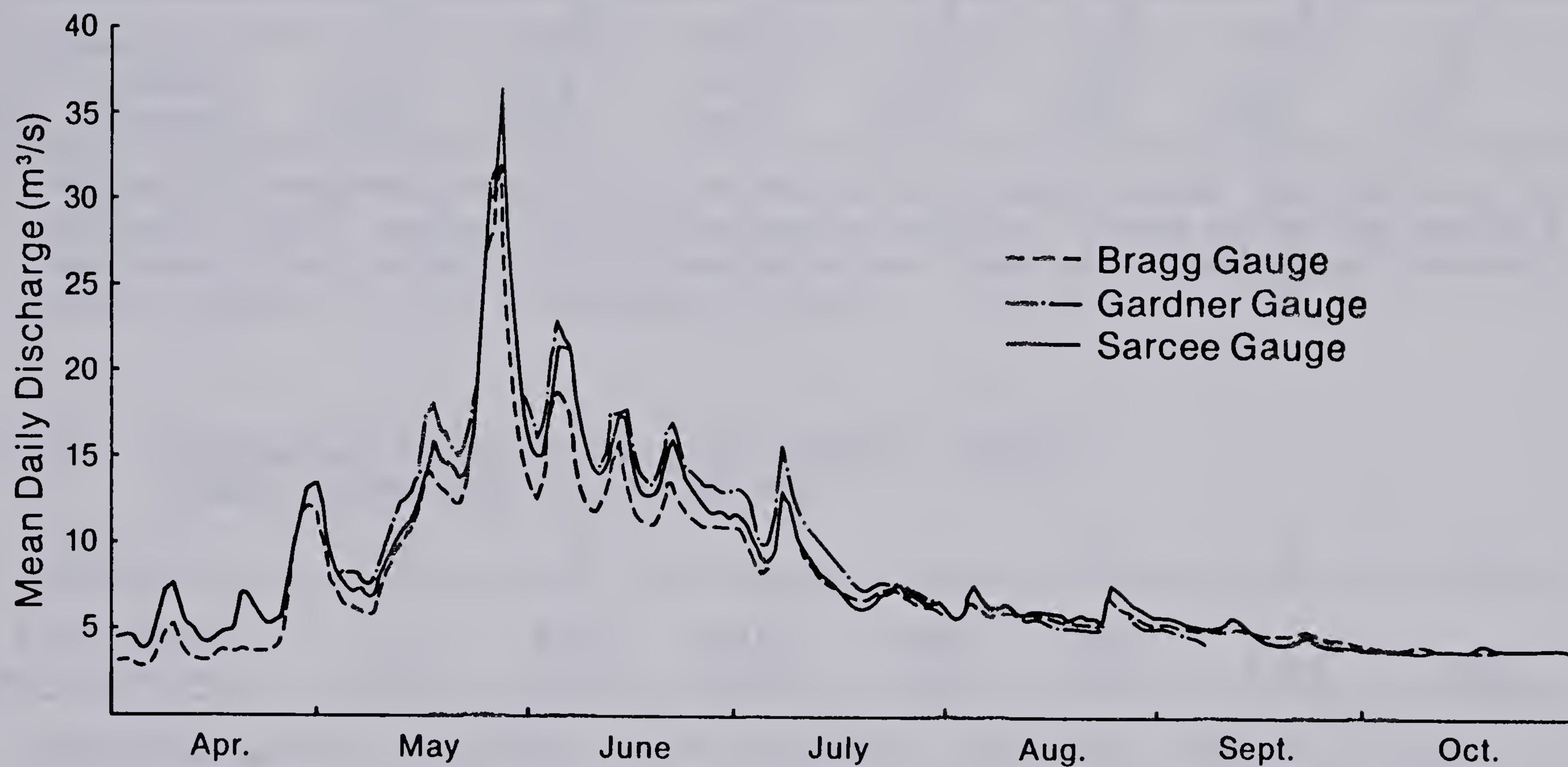

FIGURE 7 Mean daily discharge lower Elbow River basin hydrometric stations, April-October, 1979

Table 8. Estimated Floodplain Evapotranspiration Losses, Lower Elbow Basin

	A	M	J	J	A	S	O	units
pot. evapotrans.	32.3	73.6	100.8	125.0	104.9	60.5	32.3	mm
Bragg - Gardner	525	1180	1638	2031	1705	983	525	da m^3
Gardner - Sarcee	835	1877	2607	3233	2713	1565	835	da m^3

N.B.: Potential evapotranspiration calculated as the normal for Calgary by Laycock (1957) using the Thornthwaite method. This value is taken to represent the actual evapotranspiration from the floodplain because water supply is not a limiting factor.

Table 9. Estimated Gross Monthly Discharge, Lower Elbow Subbasins, 1979 (da m^3)

Reach Month	A	M	J	J	A	S	O
Bragg - Gardner		8594	9217	5298	1448		
Gardner - Sarcee	4785	6099	7867	3942	4033	1452	702

N.B.: Subbasin gross discharge = net discharge (subbasin output - input from upstream) plus evaporative losses from the river reach plus evapotranspiration losses from the floodplain.

actually increases downstream (Table 9). This increase in discharge is thought to ultimately come from the upper basin.

Direct runoff from precipitation played a minor role in the water balance of the Sarcee sub - basin in 1979 with the exception of the spring runoff period (Table 10). These results concur with the findings of Laycock (1957) who suggested that the lower Elbow River basin area does not receive sufficient precipitation in a normal year for more than minor amounts of runoff to occur.

Meyboom (1961:26) suggests that bedrock aquifers are recharged by local precipitation. The water balance studies of Laycock (1957) suggest that little of the precipitation is available for recharge because of the high evapotranspiration. The water balance approach taken here suggests that groundwater recharge occurred only in the spring of 1979 (Table 11). This supports Laycock's (1957) analysis.

When Table 6 and 11 are examined together it becomes apparent that, with the exception of the spring periods, the increase in discharge downstream from Gardener to Sarcee gauge must have an exclusively exotic source, at least in 1979. The question which must be addressed is where does this groundwater come from?

Several further lines of evidence suggest that most of the groundwater discharge in the plains area of the Elbow River basin is from the foothills and mountains to the west. Hitchon (1969:194) states that the foothills belt of the Rocky Mountains is the ultimate source of the regional flow system in Western Canada and that all relatively minor hills and valleys are local recharge and discharge areas respectively. Further, Swanson (1970) identified an interbasin flow system, of the type described in the Elbow River basin, in the Porcupine Hills, which are foothills immediately south of the Elbow River basin. Toth (1966) describes a plains area north of the Elbow River basin as having areas of downward flow (26%), flow nearly parallel to the topographic surface (42%) and areas of upward flow (32%). Pedological evidence suggests this is probably the case in the Elbow River basin.

MacMillan (1980) has described the soils of the foothills and plains area of the Elbow River basin. The principal soils of the foothills are luvisols which form where downward movement of moisture is great. In the Gardener sub - basin area soil evidence shows that there are areas of recharge and areas where the soil is neither strongly leached nor salinized. This indicates that the soil is forming under the influence of water derived from local groundwater discharge, or a very slow recharge into a high

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'(cont'd)net discharge (sub - basin output minus the input from upstream) plus atmospheric losses from the river reach and the floodplain.

Table 10. Estimated Direct Runoff, Lower Elbow Subbasins, Spring and Summer Months, 1979.

Subbasin	April	May	June	July	Aug	Sept	Units
Gardner		1.78	4.70	0.25	0		mm
Sarcee	6.99	17.78	5.08	7.62	0	0	mm

Table 11. Estimated groundwater recharge, lower Elbow sub-basin spring and summer months, 1979

Subbasin	April	May	June	July	Aug.	Sept	Units
Gardner		0.0	0.0	0.0	0.0		mm
Sarcee	5.5	3.0	0.0	0.0	0.0	0.0	mm
	1700	927	0.0	0.0	0.0	0.0	da m^3

N.B.: recharge estimated from precipitation minus direct runoff (estimated by the Soil Conservation Service runoff curve approach, Chow, 1964) minus evapotranspiration (Laycock, 1957).

water table. The higher portions of the Sarcee sub - basin have soils which developed under a balance of infiltration and evapotranspiration, or with a minor net infiltration (MacMillan, 1980). The intervening basin areas may function as aquacludes maintaining water near the surface. An isolated occurrence of solonetic soil was noted, but in general there is no morphological indication that the basin areas are affected by groundwater apart from the area near Glenmore Reservoir where net upward movement, salt accumulation and regional groundwater discharge occur. In the river valley itself, soils are largely orthic humic regogols, which are calcareous to the surface. This suggests that a net upward water movement occurs (MacMillan, 1980).

The chemical composition of the surface water (Table 5) and groundwater systems changes downstream. Ozoray and Barnes (1978:26) discuss general flow system chemical composition in the Calgary area and state: "Consequently, along the flowpath of the groundwater system, the quantity of the total dissolved solids ... usually keeps on increasing while the chemical character changes from calcium bicarbonate toward calcium-magnesium bicarbonate. If the water remains underground for a longer period, as is the case for a longer local or an intermediate flow system, the change continues toward sodium bicarbonate; if sulfate is available, the water becomes increasingly sulfatic." The specific conductance, filterable residue concentration and magnesium and sulfate concentration of river water increase downstream through the Gardener and Sarcee sub - basins during spring and summer (Table 5; Allessio & Van Dyl, 1973:A1-13).

DOWNSTREAM DECREASES IN DISCHARGE

Throughout the spring and summer major losses in discharge were reported for the Gardener to Sarcee reach. In addition, in late summer and fall a downstream decrease in discharge occurs between Bragg and Gardener gauge (Table 3).

The decrease in discharge downstream is attributed to a combination of a decrease in groundwater input with time and to atmospheric losses.

The groundwater contribution to the sub - basin may be estimated using a continuity equation approach:

$$Qgw = Qn + A \pm S - P$$

where: Qgw is the groundwater input
Qn is the net surface water discharge of the sub - basin (outflow minus inflow from upstream)
A represents atmospheric losses (E, evaporation from the river reach and ET, evapotranspiration from the floodplain)
S represents changes in soil moisture storage
P is the precipitation input

The net discharge of the sub - basin has been accurately measured. Evaporation losses have been estimated by the Meyer (1947) equation, following Schulz (1973). Evapotranspiration losses from the floodplain have been estimated from potential evapotranspiration normals (Laycock, 1957). Given that the floodplain is well vegetated, and has a high water table (MacMillan, 1980), potential evapotranspiration probably represents actual evapotranspiration (Freeze and Cherry, 1979:207).

Three components of the equation remain: (a) the groundwater discharge, (b) the precipitation storm runoff and, (c) base flow from soil moisture recharge following precipitation. The groundwater discharge can be deduced if the other two factors are estimated.

To estimate the storm runoff from rainfall the Soil Conservation Service runoff curve number approach was used (Ogrosky and Mockus, 1964:21-11 to 21-29). Soils were considered to be in good hydrologic condition with a low to moderate average infiltration rate (CN = 77.5; Ogrosky and Mockus, 1964:21-27). Storm runoff was calculated for each rain day of the summer using data from Sibbald Ranger Station which is just north west of the Gardener sub - basin. Derived runoff volumes were subtracted from the gross monthly discharges (Table 9). Baseflow derived from rainfall was found to be negligible because of the high evapotranspiration from the sub - basin (Table 11).

The results obtained agree well with the hydrologic data from Gardener gauge. For example, in May 1979 the estimated groundwater discharge for the Gardener sub - basin was 3.16 m^3/sec whereas the average daily peak instantaneous discharge at Gardener gauge was 3.15 m^3/sec higher for the mean of the daily instantaneous low flows. The groundwater discharge estimates therefore appear to be an acceptable base for discussion of decreasing discharge downstream.

The discharge of groundwater from the Gardener sub - basin decreases dramatically over the summer so that in August atmospheric losses from the floodplain and the river reach exceed the groundwater input (Table 6, 7 and 8). Hence, surface water discharge decreases downstream from Bragg to Gardener gauge.

The decrease in the groundwater contribution with time is explained by transient regional groundwater flows. Freeze and Cherry (1979:208-209) state "...the transient rate of groundwater discharge provides a measure of the baseflow hydrograph of the stream. Increased baseflow is the result of increased hydraulic gradients in the saturated zone near the stream, and, ... this is itself a consequence of increased up-basin gradients created by a water-table rise. The time lag between a surface-infiltration event and an increase in stream baseflow is therefore directly related to the time required for an infiltration event to induce a widespread water table rise." Precipitation, snow melt and

atmospheric losses are highly time dependent.

CONCLUSIONS

The discharge of the 1210 km^2 Elbow River basin above Glenmore Reservoir is dominated by runoff from the foothills and mountains which represent the upper two thirds of the basin. However, subtleties in the plains area hydrology are evident when the sub - basin runoff is examined separately from the runoff inputs from upstream.

Increases in discharge through the plains reach in winter and early spring are attributed to snow melt in the lower basin. Later in spring, when the upper basin spring runoff occurs, additional runoff in the lower basin occurs as the result of transient, rapid groundwater inputs. Water chemistry, soils, hydrologic, meteorologic and hydrogeological evidence suggest that the groundwater is derived from a shallow flowpath which emerges near the foothills - plains boundary. These groundwater inputs occur for a limited period in spring and early summer, which corresponds to the main runoff generating period of the upper basin streams.

Decreases in discharge downstream are attributed to substantial evapotranspiration losses from the lower basin floodplain.

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